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CONTENTS

	Pages
Distribution of radioactivity in granites and possible relation to helium age measurements. By Patrick M. Hurley	1–8
Lava flows in the Grand Canyon of the Colorado River, Arizona. By John H. Maxson	9–16
Geologic cross section of the Cordillera de los Andes at about parallel 33° L. S. (Argentina-Chile). By Félix Gonzalez Bonorino	17–26
Correlation of the Cretaceous-Tertiary transition in Saskatchewan and Alberta, Canada. By Loris S. Russell	27–42
Structure and petrology of the Lovewell Mountain quadrangle, New Hampshire. By Milton T. Heald	43–90

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

Moraine and Shoreline Relationships at Mono Lake, California. By William C. Putnam

Pediment Concept Applied to Shinarump and Similar Conglomerates. By Wm. Lee Stokes

Physiographic Features of the Panamint Range, California. By John H. Maxson

Gold and Silver Content of Some Trees and Horsetails in British Columbia. By H. V. Warren and R. E. Delavault

Recent Analyses for Age by Lead Ratios. By O. B. Muench

Storm King Granite at Bear Mountain, New York. By Kurt E. Lowe

Depressions of the Northern Portion of the Southern High Plains, Eastern New Mexico. By Sheldon Judson

Volcanic Geology of Toowa Valley, Southern Sierra Nevada, California. By Robert W. Webb

Ore Mineralization in the Homestake Gold Mine, Lead, South Dakota. By James A. Noble

Contact-Metamorphic Effects of Triassic Diabase at Safe Harbor, Pennsylvania. By Randolph W. Chapman

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DISTRIBUTION OF RADIOACTIVITY IN GRANITES AND POSSIBLE RELATION TO HELIUM AGE MEASUREMENT*

By Patrick M. Hurley

CONTENTS

	Page
Abstract	1
Introduction	1
Alpha particle analysis	2
Leaching tests	2
Helium retention and age measurement	5
References cited	7

ILLUSTRATIONS

Figure	Page
1. Comparison of alpha pulse heights	2

Abstract

Granulated samples of acidic igneous rocks have been found to give a much higher rate of emission of alpha particles than corresponds to the known total contents of uranium and thorium. Abnormally large residual ranges of the alpha particles suggest a surficial distribution of the radioactive elements on the surfaces of the granules in the form of secondary mineral coatings. Most of the alpha-particle activity can be removed by dilute hydrochloric acid, leaving an activity corresponding to the low content of uranium and thorium found in sandstones and arkoses, in which the essential mineral grains of granites have been rounded and cleaned by attrition. The proportion of surficial activity appears to be less for granite samples taken from considerable depth below the surface than for samples from the zone of surface-water saturation. Losses of activity by acid treatment appear to be related to the degree of retentivity of helium in different igneous rock types. Determination of the helium age ratio before and after acid treatment suggests that the essential rock minerals contain as much helium as they should commensurate with their age and with the quantity of radioelements uniformly distributed within them, and not removable by acid. The low ratio of helium to radioactivity for the rock as a whole may be due either to loss of helium from highly radioactive areas that are easily affected by acid and thus also probably by ground waters, or to actual supergene enrichment of radioelements late in the history of the rock.

* This work has been supported by the Office of Naval Research in a laboratory constructed under a grant from The Penrose Bequest of The Geological Society of America.

Introduction

In an earlier paper (Nogami and Hurley, 1948), an analysis was made of the absorption of alpha particles in a solid source. This analysis was made for the hypothetical case of radioactive atoms distributed uniformly throughout a homogeneous material. It was pointed out that absorption corrections for the counting of alpha particles from various minerals and mineral aggregates will not apply to materials wherein the radioelements are contained in crystalline structures within the aggregate for which the absorption corrections are widely different from the aggregate as a whole. As demonstrated here the alpha activity from thick sources of many igneous rocks cannot be used directly to compute the total amount present because the activity seems to be related to surfaces of the comminuted material, and to previous exposure of the rock to ground waters.

Observations of pleochroic halos, and direct measurements (Keevil, Larsen, and Wank, 1944; Poole and Bremner, 1949), have shown that the radioactive elements occur in granites largely in scattered concentrations which seem to be related to certain accessory minerals. These concentrations appear to be high relative to the content of the radioelements dispersed throughout the essential minerals.

Consider an aggregate of ground quartz and feldspar and within this matrix little crystals that contain all the radioelements in the aggregate as a whole. Even if the sample were not ground to the dimensions of the small radioactive crystals, the overall absorption correction

for the alpha-particle counting must lie between the absorption correction for the pure quartz-feldspar matrix and the absorption correction for a source made up entirely of the radioactive crystals. In nearly all cases the rate of emission of alpha particles from granitic rocks is several times the rate predicted from the total uranium and thorium content and the absorption correction for the above hypothetical distribution. Furthermore, the residual ranges of the alpha particles after they leave the surface of the sample indicates a considerably larger proportion of the radioactive atoms on exposed surfaces of the granulated material than would be the case for the hypothetical aggregate of small radioactive crystals scattered throughout a less radioactive matrix.

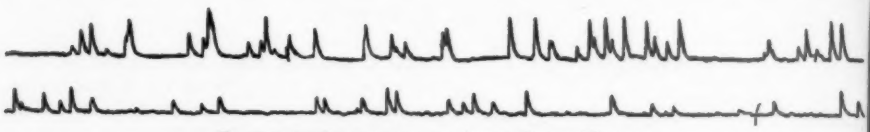

FIGURE 1.—COMPARISON OF ALPHA PULSE HEIGHTS
Upper record, granulated granite; lower record, clean arkose.

ALPHA PARTICLE ANALYSIS

For an alpha particle originating beneath a solid surface and emerging from the solid into a gas, the total ionization of the gas is given by the area under the Bragg curve for the remaining energy of the alpha particle. When an ionization chamber is used that permits the alpha particle to travel its full range in the gas, and the charges on the ions are collected and fed into a vibrating reed electrometer, the height of the "pulse" on the recording instrument (Fig. 1) is proportional to the remaining energy of the alpha particle. An analysis of these pulse heights gives the distribution of the radioactive elements within and on the surface of the solid source. If all the radioactive atoms in the uranium and thorium series were present in equilibrium amounts and were situated all on the surface of the solid source, the distribution of pulse heights would correspond to the discrete energies of the 22 alpha particles associated with these atoms. If all the radioactive atoms were distributed uniformly throughout the solid medium, this distribution of pulse heights would not be observed, and the mean height would be reduced considerably.

An apparatus for the analysis of alpha-particle ranges was not available, so that visual counts were made on the recorded pulse heights from various samples. It was found that for chemical precipitates, certain mineral samples that appear to have radioactive atoms distributed throughout their structures, and for "leached" samples to be discussed later, 95 per cent of the pulse heights at a particular instrumental sensitivity were less than 5.6 mm. Therefore, the ratio of the number of pulses exceeding 5.6 mm. in height to the total number of pulses counted can be regarded as a measure of the "superficiality" of the radioelements in the source. Figure 1 shows a comparison of a typical record for granulated granite with a record for a sample of arkose that has had the superficial activity removed from the grains. The arkose record is typical of some of the magnetite samples that appear to show correct helium ages.

It was observed that samples of granitic rocks gave the highest pulse-height ratio. Two or three of the magnetite samples showing anomalous helium age ratios also exhibited high pulse-height ratios. It appeared, therefore, that a significant proportion of the radioelements exists in surface coatings on the granules of the granite samples, rather than entirely in conjunction with essential and accessory minerals as usually supposed.

LEACHING TESTS

In order to investigate further the surficial occurrence of the radioelements a number of samples of granite were crushed to various degrees of fineness and treated with different reagents. Dilute hydrochloric acid removed almost all the uranium and/or thorium in most of the granitic samples, leaving what appeared

to be otherwise untouched grains of the essential rock minerals with very low activity. The activity of the sample was generally reduced by a factor of 5 to 10, by this treatment.

When a sample of granite is crushed, it is expected that it will break along intergranular boundaries and along surfaces within the essential minerals which are structurally weak. These surfaces may be fractures resulting from stresses in the cooling of the igneous bodies, lineages, or lines of weakness caused by the development of secondary minerals which may be either deuteric or supergene. The activity distributed along the surfaces may be in the form of scattered, minute crystals, or more uniform coatings of secondary minerals like some of the stable phosphates such as vivianite or torbernite. The thickness of the activity-carrying material must be less than the range of alpha particles in this material, otherwise the pulse-height distribution would be similar to that in Figure 1 for arkose in which the activity was homogeneously distributed throughout the mass.

It has been observed by several investigators that a large percentage of the radioactive gases radon and thoron generally escapes from pulverized samples of granite. The recoil energies imparted to the atoms of these gases when they are formed are much too low to drive them out of a crystal structure even when they occur within a few atomic layers of the surface. Possibly the material in which the parent elements of these gases occurs is so disordered, or the parent elements occur so superficially, that the atoms of radon and thoron escape readily by thermal agitation alone. This observation thus presents further evidence of extremely thin coatings of radioelements, although mechanical disruption of the crystal structure by the alpha particles may permit the migration of radon and thoron out of accessory minerals with a high content of radioelements.

It is important to take precautions against the counting of escaped radon and thoron atoms. The ionization chamber was continuously flushed with a strong stream of nitrogen to remove escaping radon and thoron. That these elements were effectively removed and not counted was demonstrated by the lack of any growth of activity which would have been observed if radon and/or thoron had escaped and disintegrated in the ionization chamber.

Measurements of the alpha-particle activity of samples of Quincy granite derived from a single specimen that was crushed and screened into four sized fractions are listed in Table 1. After several hours' treatment with dilute hydrochloric acid, the activity could not apparently be further reduced. The residual activity was the same in all four sized fractions.

TABLE 1.—MEASUREMENTS OF ALPHA-PARTICLE ACTIVITY OF QUINCY GRANITE, GROUND AND SCREENED INTO FOUR SIZED FRACTIONS

Successive Treatments	Fraction 1 420–840 microns	Fraction 2 250–420 microns	Fraction 3 125–250 microns	Fraction 4 −125 microns
1. Original sample...	9.6	7.0	7.7	10.8
2. After cold water wash.....	9.6	8.5	7.7	11.6
3. After 24 hrs. in hot water.....	9.1	7.7	7.7	7.9
4. After 2 hrs. in hot 1:5 HCl.....	1.7	1.5	1.3	1.3
5. After 16 more hrs. in hot 1:5 HCl.	1.0	1.0	1.0	1.1
6. After 16 hrs. in hot aqua regia.....	1.0			
Per cent Pulse heights exceeding 5.6 mm.....				
Original sample....	22.	17.	24.	23.
After acid leach..	4.2	4.4	4.0	4.2

Activity given in units of alpha pulses per cm^2 of source per hr. from thick granular source in a 2 π-geometry counter, with all alphas recorded that have a residual range after emergence from the source of more than 0.5 air cms.

The pulse-height ratios for the samples before and after acid treatment are also shown in Table 1. The proportion of pulses exceeding 5.6 mm. for the original untreated sample averaged 22 per cent for the four fractions, indicating a concentration of the radioelements on the surfaces of grains for all the sizes. After acid treatment, the ratio dropped to an average of 4 per cent, which is about the value found for supposedly homogeneous materials.

A rough test was made on the amount of activity in the hydrochloric acid solution, after

treating a sample of rock. It was found that an amount of uranium and/or thorium had gone into the solution that would account for all the activity lost by the rock sample.

The content of uranium and thorium in the Quincy granite has been measured several times in the past. Separate measurements of radon and thoron by Keevil (1938) for several specimens gave average values of 0.99×10^{-12} gm./gm. radium and 8.95×10^{-6} gm./gm. thorium. Goodman and Evans, and L. F. Curtiss have checked the value for radium closely. Urry (1948), in an experiment in which he added equivalent amounts of uranium series, thorium series, and potassium to a bulk sample of the granite, and thereby found the gamma radiation to be almost exactly doubled, verified these values independently. The counting rate for the experimental conditions described in the footnote to Table 1 is given by:

$$E(\alpha s/cm^2.hr) = 915\ \mu d\ (Ra + .105\ Th)$$

where Ra is in units of 10^{-12} gm./gm. and Th in units of 10^{-6} gm./gm., and both series in radioactive equilibrium. The absorption term μd is about .0016 for Quincy granite. The calculated counting rate, E, for a homogeneous distribution of radioactive atoms in the Quincy granite with the amounts of radium and thorium cited above, is 2.8 $\alpha/cm^2/hr$. Whereas the measured activity was higher than this figure before acid treatment, and lower after acid treatment, it appears that the activity was not homogeneously distributed through the sample, but concentrated on the surfaces of the granules at depths less than the alpha-particle ranges, and subject to removal by acid.

Table 2.—Reduction in the Alpha-Particle Activity of Granulated Samples of Diabase and Granite, by Acid Leach

Sample	Original Activity	Activity after HCl Leach	Ratio Final/ orig. Act.
	$\alpha/cm^2/hr.$	$\alpha/cm^2/hr.$	
327 Palisade diabase	.68	.27	.40
329 Palisade diabase	.68	.43	.73
1210 Logan Sill diabase	.76	.31	.41
1207 Logan Sill diabase	2.0	.34	.17
Average ratio for diabases			.43
304 Creighton granite	15.7	.73	.05
303 Sudbury micropegmatite	3.4	1.14	.34
1082 Keene, N. H., granite	5.5	.89	.16
1355 Quincy granite	8.8	2.0	.23
1357 Quincy granite	11.1	.86	.08
1365 Quincy granite	17.2	2.7	.16
1366 Quincy granite	8.8	1.9	.22
1368 Quincy granite	9.0	1.8	.20
Average ratio for granites			0.18

The coarse sample in Table 1 was finely ground after the aqua regia treatment, and remeasured to see if the activity would rise after fresh interior parts were exposed. It showed an activity of 0.8 $\alpha/cm^2/hr.$, even less than the activity before grinding.

Tests on sand grains derived from granitic rocks that have been rounded and cleaned by attrition have indicated that the amounts of uranium and thorium incorporated uniformly in the essential minerals are generally equivalent to an activity of less than 1.0 $\alpha/cm^2/hr.$ Such an activity is roughly the same as that found to remain in granular samples that have been subjected to dilute acid treatment.

Other tests, similar to those in Table 1, have indicated that little or no activity is removed from the rock samples by solutions containing Na_2CO_3, $NaHCO_3$, or $\frac{1}{4}$ normal acetic acid, although in the last case a slight reduction in activity might have been followed by further solution of the radioelements over a prolonged period.

The effect of acid treatment on a few specimens of diabasic and granitic rocks is compared in Table 2. The ratio of "insoluble" activity to the original activity is higher for the mafic rocks than for the granites. These ratios are strikingly similar to the "helium retentivities" of these respective rock types as will be discussed below.

To test the possiblity that the uranium or thorium may be in part supergene, three samples of granite from the deeper sections of the Adams Tunnel in Colorado, and one sample from 4000-foot depth in the Butte mining district were studied (Table 3). These samples were much tougher than the granites listed in Table 2, and breakage occurred across the grains rather than around grain boundaries. The reduction

in activity by acid treatment was much less (50 per cent) for these deep-seated granites than for the surface granites (80 per cent). Furthermore, the percentage of pulse heights exceeding the criterion for uniform distribution of radioelements averaged only 6 per cent for the deep-seated samples, as opposed to 20 per cent for all the near-surface granites listed in Tables 1 and 2. These observations point to differences between samples taken from the zone of water saturation near the surface and samples taken from depths at which there was probably little or no water. If it is true that supergene alterations have affected these changes in igneous rocks, the cause of low helium ages in igneous rocks may be due to modifications of the radioactivity late in the rock's history, rather than to losses of helium.

TABLE 3.—PULSE-HEIGHT RATIOS AND EFFECT OF ACID LEACH ON SAMPLES OF GRANITE FROM 2500–4000 FEET DEPTH

Sample	Original activity	Per cent pulses exceeding 5.6 mm	Activity after HCL leach	Per cent pulses exceeding 5.6 mm	Ratio Final/ original activity
	$\alpha/cm^2/hr.$		$\alpha/cm^2/hr.$		
Adam's Tunnel Colo. Sta. 213 + 60	3.8	7.4	2.1	4.0	.55
Adam's Tunnel Colo. Sta. 353 + 08	12.0	7.5	6.3	6.7	.53
Adam's Tunnel Colo. Sta. 429 + 31	6.3	4.4	2.7	4.5	.43
Anselmo Mine, Butte	4.3		2.0		.47
Average		6.0		4.8	0.51

HELIUM RETENTION AND AGE MEASUREMENT

The ratio of $\frac{\text{helium}}{\text{uranium + thorium}}$ is theoretically a measure of the geologic age of a rock or mineral. This ratio for granites is found usually to be only about one-fifth of what it should be for the true age of the rock (Keevil, 1941). The ratio in the case of mafic rocks, particularly for diabase dikes, has been found closer to two-fifths. Most minerals separated from these rocks have shown similar low age ratios, but magnetite appears to give approximately the correct age in a majority of the analyses (Hurley and Goodman, 1943). Low helium ratios obtained for igneous rocks have been commonly attributed to a leakage of helium from within the essential minerals. It is known that the helium escapes from minerals containing much uranium or thorium, but here the high alpha activity often shatters the crystal structure. The investigations of the leaching effect described above appear to throw some light on the question of the escape of helium from silicate minerals in which the shattering effect of the alpha particles is negligible.

The helium in the Quincy granite has been measured several times by Keevil (1938) and by the author. There has been good agreement, and it appears that the helium content of Quincy granite in the vicinity of the Swingle Quarry is fairly constant and amounts to about 4×10^{-5} cc/gm. Keevil (1938) found for the whole rock a radium and thorium content of 0.99×10^{-12} and 8.95×10^{-6} gm/gm respectively. Helium is given off by the uranium series at a different rate from that by an equal weight of thorium series, so an "Activity Index," I, is used which reports the uranium and thorium series present in terms of ($1.04\ Ra \times 10^{-12}$ gm/gm + $.0886\ Th \times 10^{-6}$ gm/gm). If this index is used the disproportionate contributions of helium by the two series are corrected for, and at the same time the production of helium is given in terms of atoms per mg. of rock per hour (Evans and Goodman, 1944). This figure is not quite the same as that given earlier for the emission of alpha particles from a square centimeter of the rock surface per hour, and should not be confused with it. According to Keevil's measurements, therefore, the Quincy granite has an Activity Index of 1.82 α/mg/hr. From the simple age equation of:

$$\text{Age} = 30.7 \frac{(He \times 10^{-5}\ \text{cc./gm.})}{I} \text{ million years}$$

the age ratio of the Quincy granite is 68 million years. From geological evidence the granite is believed to be between Devonian and Permian in age, or older than 200 million years.

A sample of the granite treated with hydrochloric acid for 20 hours gave an activity of 1.0 $\alpha/cm^3/hr$. Regrinding resulted in no change, so this residual activity arises most probably from radioelements that are homogeneously distributed. The value of I corresponding to this emission of 1.0 $\alpha/cm^3/hr$. is 0.62 $\alpha/mg/hr$. The helium content of this acid-treated sample was found to be 4×10^{-5} cc/gm indicating that the helium content was not appreciably affected by the acid treatment. The helium ratio for the acid-treated sample is thus equivalent to an age of 200 million years, much more in line with the geological evidence.

This is only a preliminary analysis of the problem, but in this single instance it appears that the helium generated in the highly radioactive surficial materials that were presumably removed by the acid wholly escaped either from the rock *in situ* or the sample during preparation, whereas the helium generated by the radioelements homogeneously dispersed throughout the essential minerals, which were not attacked by the acid, was almost wholly retained from the time of formation of the Quincy granite.

A simple measurement of the helium age ratio following acid treatment on either the granitic or mafic rocks in Table 2 might yield ages that conform with other time scales. A vindication of such a treatment is afforded by the following consideration. Diabases appear to retain about 40 per cent of the expected helium content commensurate with their age (Keevil, 1941), but according to Table 2 the homogeneously dispersed activity is almost the same percentage of the total activity (43 per cent). Similarly the "helium retentivity" of granites is about 20 per cent, and the residual nonsurficial activity according to Table 2 about 18 per cent of the total activity. Further studies are necessary to establish the fact that the acid treatment removes just that surficial portion of the total activity for which the generated helium has been lost.

It is significant that the minerals that have shown reasonable helium ages—magnetite, pyrite, galena—have all been of hydrothermal origin. There is no reason to suppose that these minerals were ever subjected to solutions or end-phase liquids containing higher concentrations of the radioelements than the solutions from which the minerals grew; whereas this is a real possibility with the changing liquid phase of a crystallizing magma. The magnetite samples that have yielded reasonable age ratios have been of pyrometasomatic or of low-temperature hydrothermal origin; on the other hand the group of magnetite samples showing anomalously low age ratios have been concentrates from granitic rocks, and therefore possibly contaminated on the surface by small amounts of highly radioactive material developed in the last phase of crystallization, or by a redistribution of this radioactive material by supergene waters. A sample of magnetite obtained by magnetic concentration from a granite of probably Devonian or Carboniferous age at Surry Mt. Dam, N. H., showed the following:

Activity index	= 1.74
Helium	= 2.84×10^{-5} cc./gm.
Age	= 50 M. Y.

The sample was treated with 1:1 hydrochloric acid for 2 hours, and bright, clean magnetite was left after this treatment. The activity index was now only 0.125, and the helium content 1.1×10^{-5} cc./gm. indicating an age of 270 million years. This age for the granite would be quite reasonable in the present lead time scale. Both the activity and helium content were decreased by the acid treatment, but magnetite is soluble in hydrochloric acid, and a fairly thick layer was probably removed from the magnetite granules. This layer must have contained a much higher concentration of helium than the inner parts of the granules; a more highly radioactive surface layer on the magnetite grains would have produced a high concentration of helium in the adjacent subsurface layer. Removal of both layers appears to result in a helium ratio about equivalent to the correct age of the magnetite. The granite from which the magnetite was separated gave an activity index of 1.85 which is very close to that of the original magnetite. This suggests an overall, rather uniform, surficial activity of both the rock and magnetite.

Thus, magnetites which have been concen-

trated from granitic rocks do not appear to possess any lower capacity to retain helium than those of hydrothermal origin; in fact it would be strange to find any difference. The explanation appears to lie in a surficial contamination encountered in granitic magnetites but not in those of hydrothermal origin.

The fact that magnetite of hydrothermal origin shows apparently correct helium ratios may not be due, therefore, to any unusual retentivity of the magnetite structure. Most minerals, if they were free from contaminating mineralization or alterations might yield correct helium age ratios. The activity and helium content of common silicates of hydrothermal origin are almost always too low to determine without extreme precaution and lengthy measurement. Therefore nearly all measurements have been made on quartz, feldspar, pyroxene, and other minerals, separated from igneous rocks. However, it now appears that what had been measured was, to an appreciable extent, the activity of other minerals or aggregates present as surface coatings, for which the low-activity essential minerals above were merely the hosts. On the other hand, the helium measurements were probably indicative of the helium content of the low-activity essential minerals, the helium generated by the radioelements of relatively high concentration in the surface "contaminants" having been lost by leakage. It therefore appears necessary to reconsider the supposedly unfavorable "helium retentivities" of the essential minerals and to re-examine the apparently low helium ratios of the silicate minerals in the light of the possibilities of highly radioactive surface contaminants of very low helium retentivity.

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Massachusetts Institute of Technology, Cambridge 38, Mass.

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AERIAL PANORAMA OF VOLCANIC FEATURES IN THE WESTERN PART OF THE GRAND CANYON

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LAVA FLOWS IN THE GRAND CANYON OF THE COLORADO RIVER, ARIZONA

By John H. Maxson

CONTENTS

	Page
Abstract	9
Introduction	9
Location	9
Acknowledgments	10
Previous work	10
Older lavas of the plateaus	11
Canyon lavas	11
General statement	11
Source of the intracanyon flow	11
Distribution of remnants of the intracanyon flow	12
Conditions of extrusion	12
Lava cascades and dams	12
General statement	12
Toroweap cascade and dam	13
180–182 mile cascades	13
187 mile cascade	13
Queantoweap cascade and dam	13
Dam below 194½ miles?	14
Résumé of volcanic history	14
References cited	14

ILLUSTRATIONS

Figure	Page
1. Index map showing location of intracanyon lava flow in the western part of the Grand Canyon	10

Plate	Facing page
1. Aerial panorama of volcanic features in the western part of the Grand Canyon	9
2. Lower Canyon lavas at Lava Falls and remanants of lake beds downstream	12
3. Remnants of intracanyon lava flow	13
4. Volcanic features near mouth of Toroweap Valley	14
5. Volcanic features near mouth of Queantoweap Valley	15
6. Plan and profile of the western part of the Grand Canyon showing location of volcanic features	16

Abstract

Within the western part of the Grand Canyon, over a distance of more than 84 miles, from Toroweap Valley almost to the Grand Wash cliffs, flow remnants of olivine basalt occur along the Colorado river channel. The remnants in the downstream three-fourths of the stream segment individually pertain to single flows. The accordance of their upper surfaces with a gradient parallel to that of the present river and the failure to discover sources of lava in this segment suggest that the remnants belong to the same flow. In the upstream one-fourth of the stream segment several flows may be distinguished, the lowest of which is correlated with the downstream remnants on the basis of near-continuity and similarity in physiographic position. The source of this flow is in the vicinity of Lava Falls at the mouth of Toroweap Canyon where the farthest upstream remnants are found. Later flows from the esplanade of the Uinkaret Plateau locally and briefly dammed the Colorado River near the mouths of Toroweap and Queantoweap valleys. After removing the dams, the Colorado River cut downward to approximately its channel prior to the first flow. At one place below Lava Falls the river channel is still being cut in basalt, but everywhere else it is being cut a few feet to a few tens of feet below the bottoms of the lowest flow remnants. This long flow was an important event in the later geological history of the Grand Canyon and affords a striking example of the fluidity of basaltic lava, involving a descent of 700 feet in 84 miles along a water grade.

Introduction

Location

The lava flows described in this paper occur in the western part of the Grand Canyon of the Colorado River. This portion of the Grand Canyon lies in northwestern Arizona between the Shivwits and Uinkaret plateaus on the north and the Coconino Plateau on the south (Fig. 1). Volcanic peaks are found on the plateaus; Shivwits Plateau has Mt. Dellenbaugh and an extensive lava cap, the Uinkaret Plateau has Mt. Emma, Mt. Logan, Mt. Trumbull (Pl. 6), and an extensive lava cap, and the Coconino Plateau has the San Francisco Mountains and Bill Williams Mountain (Fig. 1).

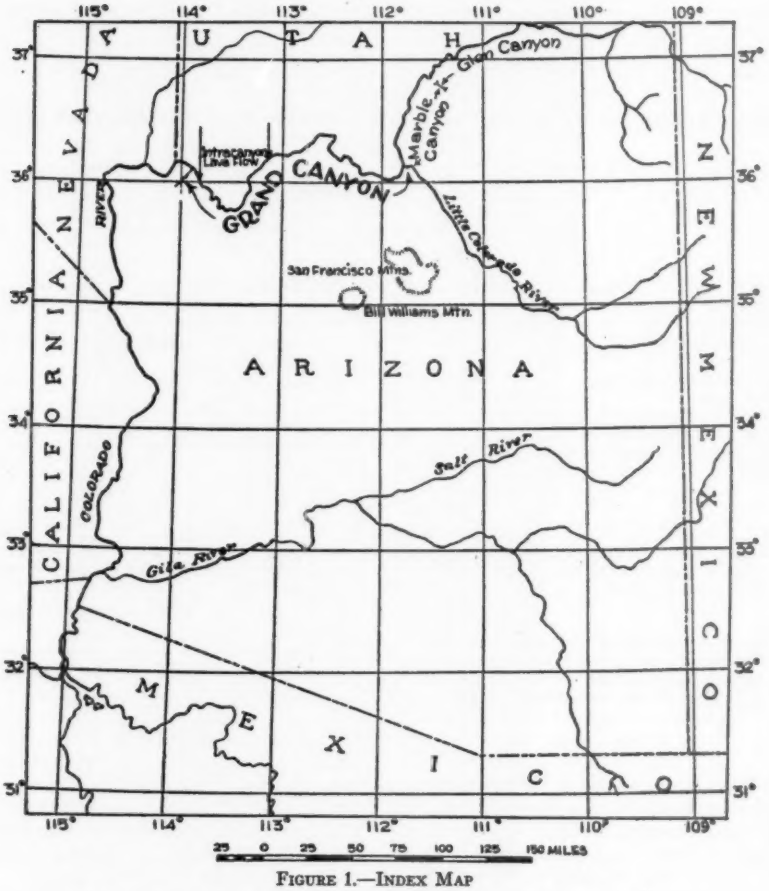

FIGURE 1.—INDEX MAP

Showing location of intracanyon lava flow in western part of the Grand Canyon

Acknowledgments

The field work on which this paper is based was done in the fall of 1937 during a boat expedition through the Grand Canyon. The study was incidental to the investigation of the Archean rocks of the Grand Canyon, a project sponsored by the Carnegie Institution of Washington through its president at that time, the late Dr. John C. Merriam. The writer wishes to express his appreciation for this support. He also wishes to thank Dr. Ian Campbell of the California Institute of Technology, who participated in the field work and who made important contributions to the observations discussed in this paper. The writer wishes to thank Mr. E. D. McKee for many valuable suggestions.

Previous Work

The first description of the volcanic phenomena in the western part of the Grand Canyon was given by Major J. W. Powell (1875, p. 94–95) after his classical river voyage in 1869. Captain C. E. Dutton (1882, p. 94), in 1879 and 1880, studied basalts of the Shivwits and Uinkaret plateaus and the lava cascades. Dr. R. C. Moore, geologist of the U. S. Geological Survey Colorado River expedition of 1923, ob-

served the position of many remnants of flows from Uinkaret Plateau into the Grand Canyon. His discussion (Moore, 1925, p. 125–171) is concerned primarily with geologic features of possible dam sites, but he discusses and illustrates some of the lavas in the vicinity of Lava Falls.

A remnant of olivine basalt 16½ miles east of Pierce's Ferry was described by Schenk (1938, p. 251) as 50 feet thick and having its base 90 feet above the low-water level of the river. Near the base of the flow he observed pillowlike structure. This remnant may be the westernmost one in the Grand Canyon; it occurs 263 miles by river below Lee's Ferry, Arizona. The writer observed this remnant in 1935 from the air (Pl. 3, fig. 3). When the locality was passed in 1937 it was inundated by Lake Mead.

McKee and Schenk (1942, p. 245–273) made a detailed study of the volcanic history near the mouth of Toroweap Valley. They subdivided the lavas within the Grand Canyon into Lower Canyon, Middle Canyon, and Upper Canyon groups. The latest study by Koons (1945, p. 151–180) establishes the volcanic sequence of the Uinkaret Plateau.

Older Lavas of the Plateaus

Extensive basaltic lava fields occur on the westernmost plateaus of the Grand Canyon—the Shivwits, the Uinkaret, and the Kanab (Pl. 6). As described by Dutton (1882, p. 109) "The ancient basalts are much eroded. Not only have their craters been demolished, but the massive floods which emanated from them have been greatly wasted." Koons (1945, p. 157–159) distinguishes flows of four stages. Stages I and II are extensively weathered and eroded. Stage III flows have been slightly weathered and eroded. The single stage IV flow on the Uinkaret Plateau is fresh in appearance. The older lavas of the plateaus belong to Stages I and II.

The older lavas of the western plateaus may be correlatives of the first volcanic episode of the San Francisco Mountains on the Coconino Plateau. Here Robinson (1913, p. 95) considers the first volcanic activity marked by widespread eruptions of basalt shortly after the development of a Pliocene peneplain.

Canyon Lavas

General Statement

The lavas of the inner canyon at the foot of Toroweap Valley have been subdivided by McKee and Schenk (1942, p. 251) into Lower, Middle, and Upper Canyon groups of successively younger age. McKee and Schenk consider the Lower Canyon group of lavas to be closely related in time and to have formed a lava dam at the mouth of Toroweap Valley. The lower two units of the Lower Canyon group (A and B flows) are described as single, massive flows. The upper two units (C and D flows) are made up of series of thin flows. As these authors point out (1942, p. 263–264) the Lower Canyon flows are nearly horizontal at the mouth of Toroweap Valley, whereas the Middle Canyon and Upper Canyon lavas approached the Colorado River from the north and poured over the front of the older lavas. McKee and Schenk traced the A and B flows for a mile along the north side of the Colorado River below Toroweap Valley. Both flows are thick and of dense olivine basalt and both have columnar jointing developed in their basal portions. One of them is the correlative of the basal flow a few miles downstream and of the remnants still farther downstream where there was but a single flow. This most extensive of the canyon lavas is termed the intracanyon flow.

Source of the Intracanyon Flow

The farthest upstream occurrence of flow remnants whose upper surfaces are approximately at stream grade is at Lava Falls on the Colorado River. This is 179 miles below Lee's Ferry as shown on sheet F of the United States Geological Survey's *Plan of the Colorado River between Lee's Ferry, Arizona and Black Canyon, Arizona-Nevada.* Future references to location of features along the river course will be made by giving the number of miles below Lee's Ferry. Only one occurrence of basalt was noted above 179 miles, Lava Pinnacle, or Vulcan's Forge, at 178 miles. This is a basalt rock about 30 feet high and 20 feet in diameter possessing a confused columnar structure. Since no probable source was observed in its vicinity, it is

believed to represent a remnant of the upstream backflow of one of the Lower Canyon group.

At 179½ miles, Lava Falls, the mouth of ancestral Toroweap Canyon, the massive, horizontal A and B flows may be observed north of the river at the base of the lava section (Pl. 2, fig. 1). There is a break between 180 and 183 miles, but from that point several miles down the river extends a single massive flow remnant (Pl. 3, fig. 1). With a few breaks in continuity this flow may be traced 10 miles farther downstream to the vicinity of 193 miles where it still has comparable thickness and structural appearance (Pl. 3, fig. 2). In view of the near-continuity and similarity in physiographic position the intracanyon flow is regarded as one of the first two flows at the mouth of ancestral Toroweap Canyon and is believed to have originated in this vicinity.

The intracanyon lava flow may have been erupted on the esplanade and flowed down ancestral Toroweap Canyon to the river. Another possibility is that fissures trending generally north-south, extending across the Colorado channel and up the old Toroweap Canyon, may have conducted the lava directly into the canyon. Opening of conduits subsequent to the A flow is shown by a number of basalt dikes which cut the flow and its subjacent river gravels.

Since basaltic lavas reached the channel from the north rim at various places and because basalt dikes have been found in the channel at 179½ miles, the question arises as to whether the flow remnants downstream from 193 miles could have originated as separate flows. The fact that all the remnants are of one rock type, olivine basalt, does not prove derivation from one source since all the lavas of the region are of this type. Of some importance is the fact that below Queantoweap Valley where several flows are present there is no thinning of the lowest flow. Thus there is no evidence of termination of one flow and assumption of its physiographic position (lowest in the channel) by another, later flow. The upper surface may be projected between flow remnants at the present gradient of the Colorado River. This physiographic evidence strongly suggests that all the remnants below 193 miles and the basal flow from 179 miles to 193 miles belong to a single flow, the intracanyon lava flow.

Distribution of Remnants of the Intracanyon Flow

The location of flow remnants and their relationship to the river profile is shown diagrammatically in Plate 6. The remnants are longer and closer together in the vicinity of the source. However, numerous remnants are found in the stretch from 64 to 78 miles below the source. The remnants are preserved because of the recency of the flow, which presumably was extruded in late Pleistocene or even early Recent time. The river is flowing close to the channel cut prior to the extrusion, as indicated by the extension of the basalt in place under water of the Colorado River at 183 miles.

The large size and thickness (50 feet) of the farthest downstream remnant at 263 miles (Pl. 3, fig. 3) suggest that the intracanyon flow actually extended a number of miles, possibly tens of miles, farther downstream.

Conditions of Extrusion

A high degree of fluidity of the basalt at time of extrusion is indicated by its flow of at least 84 miles along a stream channel. Where observed by the writer columnar structure is developed at the base of the flow overlying river gravels (Pl. 2, fig. 1; Pl. 3, fig. 2). This indicates rather slow cooling at the base and does not suggest that the flow encountered any large volume of river water. It is likely that the basalt was rapidly extruded and rapidly flowed down the channel while the river was temporarily impounded above the source.

Lava Cascades and Dams

General Statement

The lava cascades are the most spectacular recent geological feature in the western Grand Canyon district. They are very conspicuous from positions on the ground and in the air because their black lavas have flowed over the bare flats of the stripped esplanade and down the bare stepped cliffs of the canyon. In Plate 6 the lava streams from the Uinkaret Plateau are not distinguished in pattern from the older basalt sheets of the plateaus. All the cascades are from the Uinkaret Plateau north of the Colorado River. They all originate in Stage III flows (Koons, 1945).

FIGURE 1. LOWER CANYON LAVAS AT LAVA FALLS

FIGURE 2. LAKE BEDS AT 180½ MILES
Tuffs lie on river gravels which overlie angular detritus.

FIGURE 3. LAKE BEDS AT 194½ MILES
Tuffs lie on river gravels which overlie remnant of intracanyon lava flow.

LOWER CANYON LAVAS AT LAVA FALLS AND REMNANTS OF LAKE BEDS DOWNSTREAM

FIGURE 1. INTRACANYON LAVA FLOW 183–185 MILES
Continuous remnant of single flow on north side of river. On right 182 mile cascade. Lava-covered esplanade of Uinkaret Plateau in middle background. Aerial photograph by author looking northwesterly from altitude of 8000 feet.

FIGURE 2. INTRACANYON LAVA FLOW AT 189 MILES
Heavy basal flow with columnar jointing at bottom overlies stratified river gravels.

FIGURE 3. INTRACANYON FLOW AT 263 MILES
Farthest downstream remnant having thickness of 50 feet suggests that flow extended a number of miles farther down the canyon. Vertical aerial photograph by author from altitude of 10,000 feet.

REMNANTS OF INTRACANYON LAVA FLOW

Toroweap Cascade and Dam

A canyon was cut along the Toroweap fault leading from Toroweap Valley into the Grand Canyon prior to extrusion of the lavas. Parts of the walls of this ancestral Toroweap Canyon are clearly discernible (Pl. 4, fig. 1). McKee and Schenk (1942, p. 263–264) describe the filling of this canyon by Middle Canyon lavas which flowed down over the front of older lavas to the river. Like Dutton (1882, p. 95) and Moore (1925, p. 156) McKee and Schenk (1942, p. 265–266) believe that flows dammed the Colorado River at ancestral Toroweap Canyon. McKee and Schenk consider that a dam 550 feet high was formed by Lower Canyon lavas (A,B,C, and D flows) only, because they identified remnants of these flows on the south side of the Grand Canyon. They did not observe any remnants which they could correlate with the Upper and Middle Canyon lavas. The Middle Canyon lavas possessed structural evidence of flow down the north wall of the Grand Canyon and were presumed to have moved no farther than the river.

A large lava remnant on the east side of Prospect Canyon south of the Colorado River and opposite ancestral Toroweap Canyon (Pl. 4, fig. 2) suggests by its form an ancestral Prospect Canyon. It was eroded along the Toroweap fault presumably to a depth of about 1500 feet like ancestral Toroweap Canyon. The lower part of the Prospect Canyon remnant appears to fill the ancestral Prospect Canyon like the Middle Canyon lavas fill ancestral Toroweap Canyon. The upper part of the remnant is horizontal like the Upper Canyon lavas on the north rim. If the analogy in position means correlation and is not merely coincidental a high lava dam was formed subsequent to the lower lava dam of the A,B,C, and D flows. No lacustrine deposits have been reported in the Grand Canyon above Toroweap Valley. Perhaps the dam, if it existed, was so short-lived that little deposition occurred and such soft deposits as accumulated rapidly washed away.

Another lava remnant, higher than the Lower Canyon flows, exists on the south side of the river just below Lava Falls. If this and the Prospect Canyon remnant originated on the south side there would be no implication of damming by the higher lavas. They would then be remnants of cascades in and near Prospect Canyon similar to those of Toroweap Canyon on the north side. Although there are several cinder cones on the esplanade south of the river none appears to be associated with extensive lava flows. No flows seem to be connected with the Prospect Canyon remnant. However, before south-side origin can be ruled out more detailed study in the vicinity will be necessary.

180–182 Mile Cascades

Extensive cascades from the Uinkaret esplanade reached the bottom of the canyon at 180 miles and 182 miles and probably joined the lavas from Toroweap Canyon (Pl. 4, fig. 2; Pl. 3, fig. 1).

187 Mile Cascade

This small cascade (Pl. 5, fig. 1) descended a narrow valley cut along a nearly vertical fault, subsidiary to the main Hurricane fault. The block to the east has been relatively elevated approximately 200 feet. A small amount of lava reached the river channel and flowed out over the lowest Canyon flow.

Queantoweap Cascade and Dam

The Queantoweap Cascade resembles the Toroweap Cascade. Coalescing cascades from the Uinkaret esplanade flowed down an ancestral Queantoweap Canyon which was about 1000 feet deep at its mouth on the Colorado River (Pl. 5, fig. 1). The ancestral Queantoweap Canyon was eroded headward along the east branch of the Hurricane fault and cut back several miles into the relatively flat-floored Queantoweap Valley. The flows filled the narrow ancestral Queantoweap Canyon, covered the Lower Canyon flows, filled the bottom of the Grand Canyon to a depth of approximately 1000 feet, and flowed at least as far as Boulder Wash (193 miles). The upper surfaces of the flow remnants near Whitmore Wash (188 miles) slope gently downstream, and high-lying remnants occur on the east side of Boulder Wash;

remnants of high-lying flows were not observed farther down the canyon. Remnants of the Queantoweap flows occur on the south side of the Colorado River (Pl. 5, fig. 2). With a height of 1100 feet the Queantoweap dam could have backed up the waters of the Colorado River throughout the Grand Canyon (*i.e.*, as far upstream as the mouth of the Little Colorado River). However, it may have been essentially contemporaneous with an even higher dam at Toroweap. In this case the impounding would have extended upstream only to Lava Falls. Lake beds found at 180½ miles (Pl. 2, fig. 2) just below Lava Falls may have been deposited in this lake.

The conversion of the ancestral Queantoweap Canyon into a high-lying surface and the subsequent erosion of Whitmore Canyon and the new valley west of the lava cascades which feeds into it afford an interesting example of inversion of relief. It recalls Davis' prediction of inversion on the margin of the Uinkaret Plateau (Davis, 1903, p. 24).

Dam Below 194½ Miles?

The occurrence of ash beds of apparent lacustrine deposition on a pillar of columnar basalt at 194½ miles (Pl. 2, fig. 3) demonstrates that the lowest or intracanyon flow antedated the formation of a lake whose dam was below 194½ miles. Stream gravels on top of the lava pillar and under the lake beds show that the river had established its course on top of the lava before the damming downstream. No flows from the esplanade of the Shivwits Plateau to the river have been observed. The esplanade is bare of lavas. Whether the damming resulted from a flow originating within the canyon which escaped observation during the 1937 boat expedition or from landslides is not known.

The lake beds at 180½ miles and at 194½ miles are similar lithologically. Both consist of horizontal, fine-grained, olive-green tuffs. The tuffs at 180½ miles are about 50 feet above the river which is at 1650 feet elevation there. The river is at an elevation of 1550 feet at the 194½ miles location, and the tuffs are about 100 feet above the river. Thus the upstream lake beds occurrence is approximately 50 feet higher than that downstream. A dam only 100–150 feet high just below 194½ miles could have impounded the water in which both sets of lake beds were deposited. Since the tuffs are soft and easily eroded and since the beds are lithologically similar the single-lake concept is strengthened. It may have existed at the time of relatively recent cinder-cone construction when quantities of pyroclastics were carried into the river.

Résumé of Volcanic History

(1) Extrusion of basalts (older lavas of the plateaus) on stripped Mesozoic strata of western Grand Canyon region. (Late Pliocene?)

(2) Uplift and canyon erosion. Movement on Hurricane and Toroweap faults. Development of narrow esplanade by recession of Coconino sandstone cliffs over easily eroded Hermit shale. Cutting of canyon to present depth and profile (Pleistocene?)

(3) Eruption of Lower Canyon group of lavas from vents on esplanade in vicinity of old Toroweap Canyon or from fissures in Grand Canyon bottom. Damming of river. Longest or intracanyon flow extends 84 miles downstream on stream gradient losing only 700 feet in elevation.

(4) Eruption of basalt flows from vents on Emma platform of Uinkaret Plateau and the Uinkaret esplanade. Flows fill tributary canyons and cascade down Grand Canyon walls. Local damming of river, followed by breaching, and removal of bulk of the lava from the canyon. Eruption of cinder cones. (Late Pleistocene to Recent)

(5) Reduction in size of flow remnants which remained plastered on canyon walls. Re-establishment of pre-Canyon lava stream gradient throughout course covered by flow. (Recent)

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FIGURE 1. TOROWEAP CASCADE AND TOROWEAP VALLEY
Portion of side walls of ancestral Toroweap Canyon shown by dashed lines. Dotted lines show top of Redwall limestone. Toroweap fault on which right block has been raised 632 feet (McKee and Schenk) passes under lava filling. Aerial photograph by author looking north from altitude of 8000 feet.

FIGURE 2. GRAND CANYON AT TOROWEAP VALLEY
Toroweap cascade in center, 180 mile cascade left foreground, Prospect Canyon remnant middle right. Aerial photograph by author looking easterly from altitude of 8000 feet.

VOLCANIC FEATURES NEAR MOUTH OF TOROWEAP VALLEY

FIGURE 1. QUEANTOWEAP CASCADE AND QUEANTOWEAP VALLEY
Side walls of ancestral Queantoweap Canyon shown by dashed lines. 187 mile cascade right center. Basal or intracanyon flow (Pl. 3, fig. 2) near river in lower left. Aerial photograph by author looking north from altitude of 8000 feet.

FIGURE 2. INTRACANYON LAVA FLOW (emphasized by vertical lining) AND OVERLYING QUEANTOWEAP CASCADE (top shown by dashed line)
Aerial photograph by author from altitude of 7000 feet looking northeasterly.

VOLCANIC FEATURES NEAR MOUTH OF QUEANTOWEAP CANYON

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1585 Kearney St., Denver, Colo.

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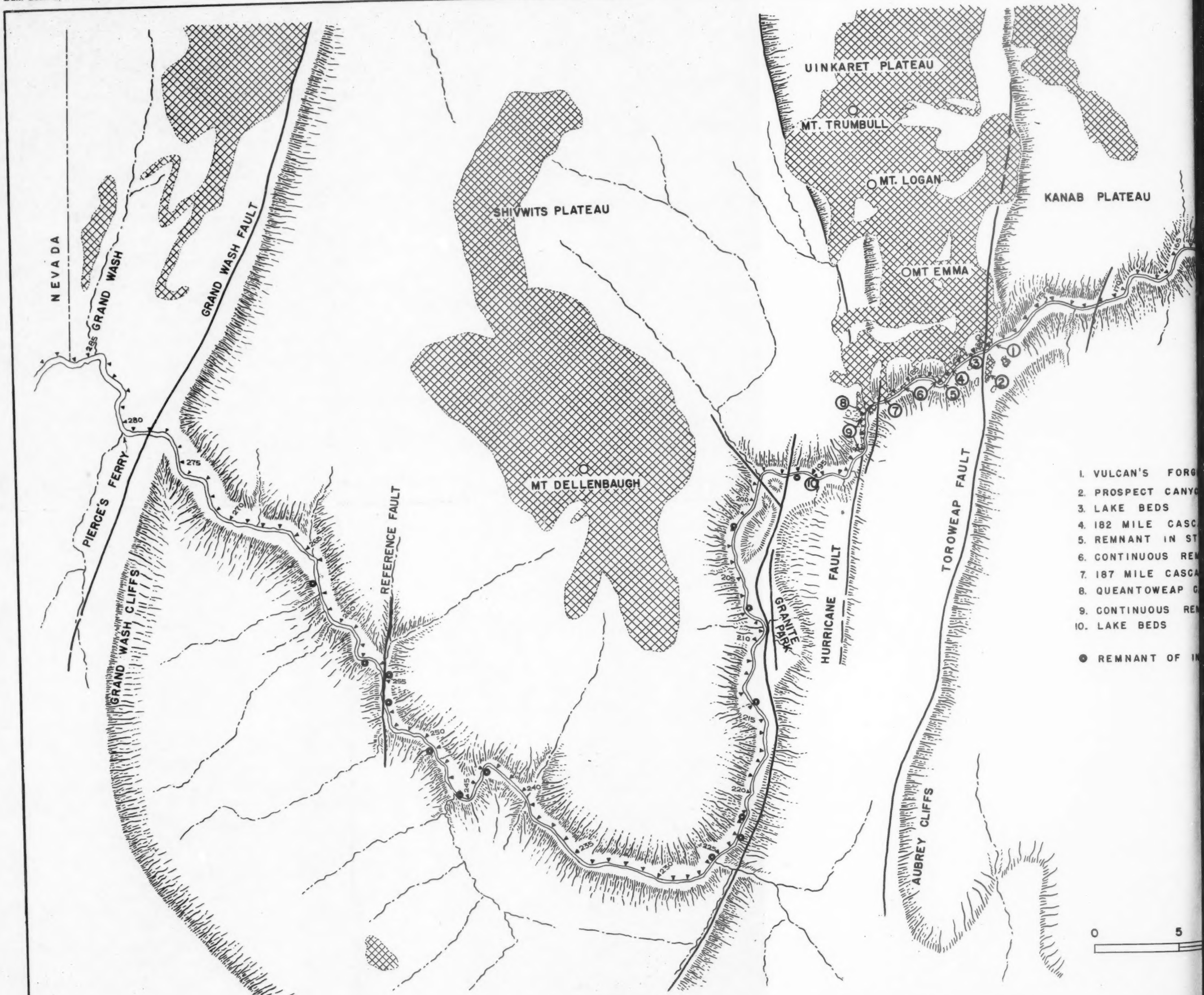


VOLCANIC FEATURES OF THE WESTERN GRAND CANYON

(DISTRIBUTION OF PLATEAU LAVAS AFTER N.H. DARTON MODIFIED)

UINKARET PLATEAU

MT. TRUMBULL

MT. LOGAN

MT EMMA

SHIVWITS PLATEAU

KANAB PLATEAU

MT DELLENBAUGH

REFERENCE FAULT

GRANITE PARK

HURRICANE FAULT

TOROWEAP FAULT

AUBREY CLIFFS

LEGEND

1. VULCAN'S FORGE
2. PROSPECT CANYON REMNANT
3. LAKE BEDS
4. 182 MILE CASCADE
5. REMNANT IN STREAM CHANNEL
6. CONTINUOUS REMNANT OF INTRACANYON FLOW
7. 187 MILE CASCADE
8. QUEANTOWEAP CASCADE
9. CONTINUOUS REMNANT OF INTRACANYON FLOW
10. LAKE BEDS

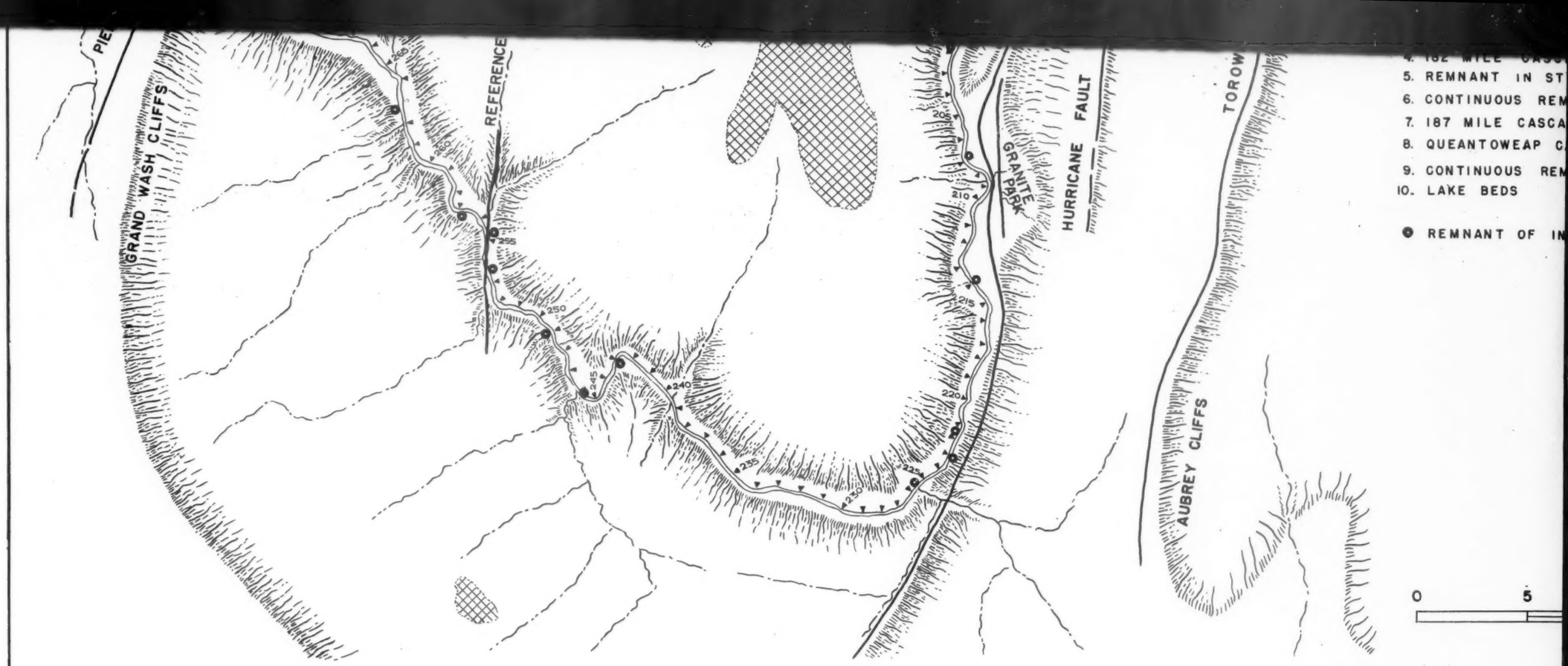
● REMNANT OF INTRACANYON FLOW

SCALE

0 5 10 15 20 MILES

VOLCANIC FEATURES OF THE WESTERN GRAND CANYON

(DISTRIBUTION OF PLATEAU LAVAS AFTER N.H. DARTON MODIFIED)



VOLCANIC FEATURES OF THE WESTERN GRAND CANYON

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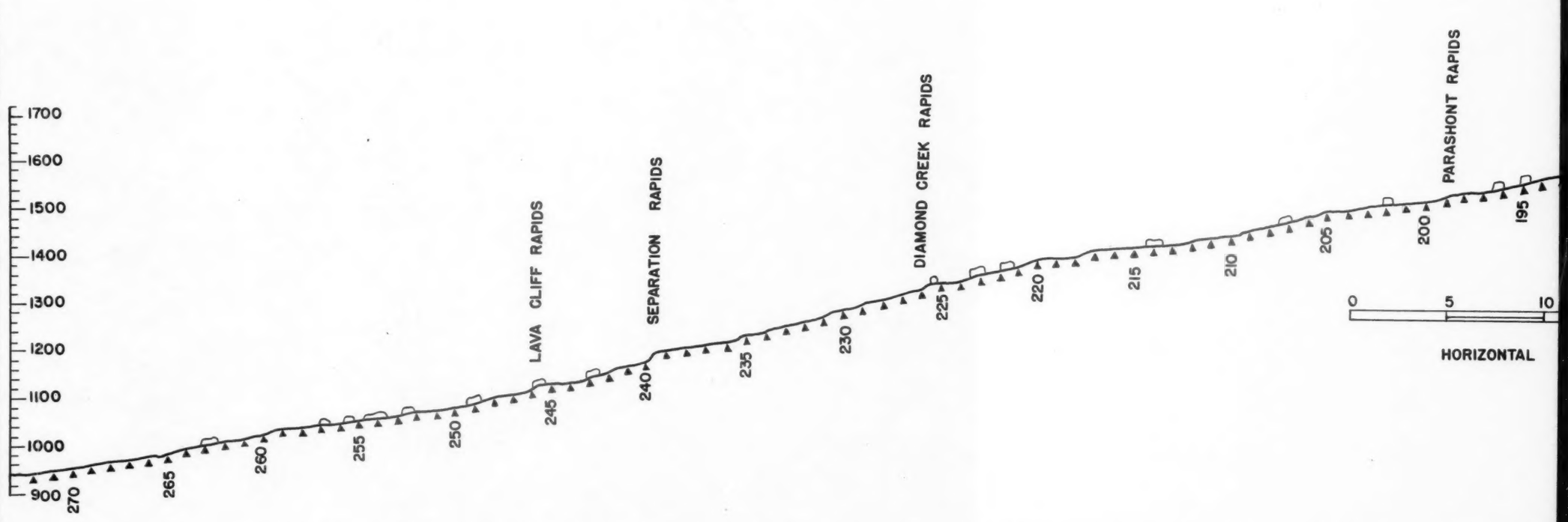
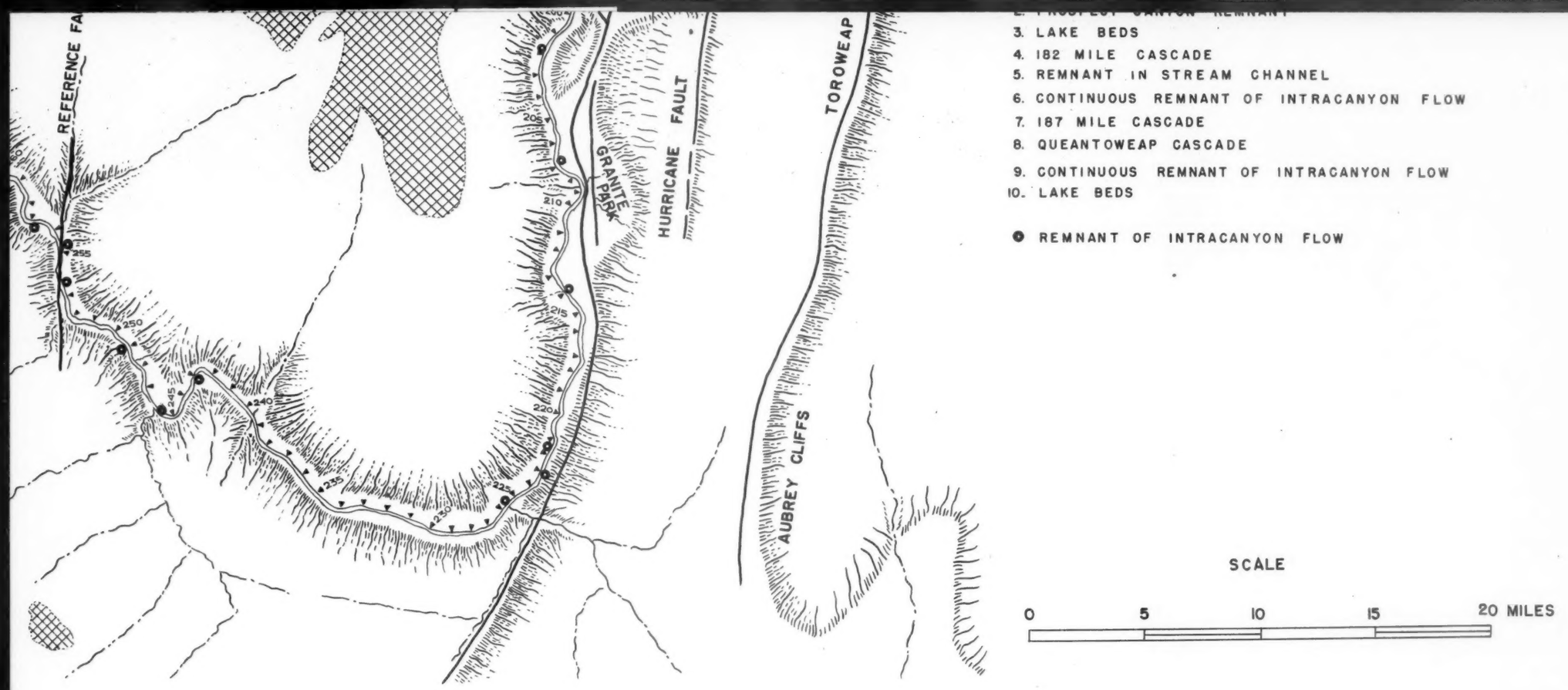


PROFILE OF COLORADO RIVER IN WESTERN PART OF THE GRAND CANYON SHOWING LOCATION OF BASALTIC LAVA

PLAN AND PROFILE OF THE WESTERN PART OF THE GRAND CANYON

Showing location of volcanic features



VOLCANIC FEATURES OF THE WESTERN GRAND CANYON

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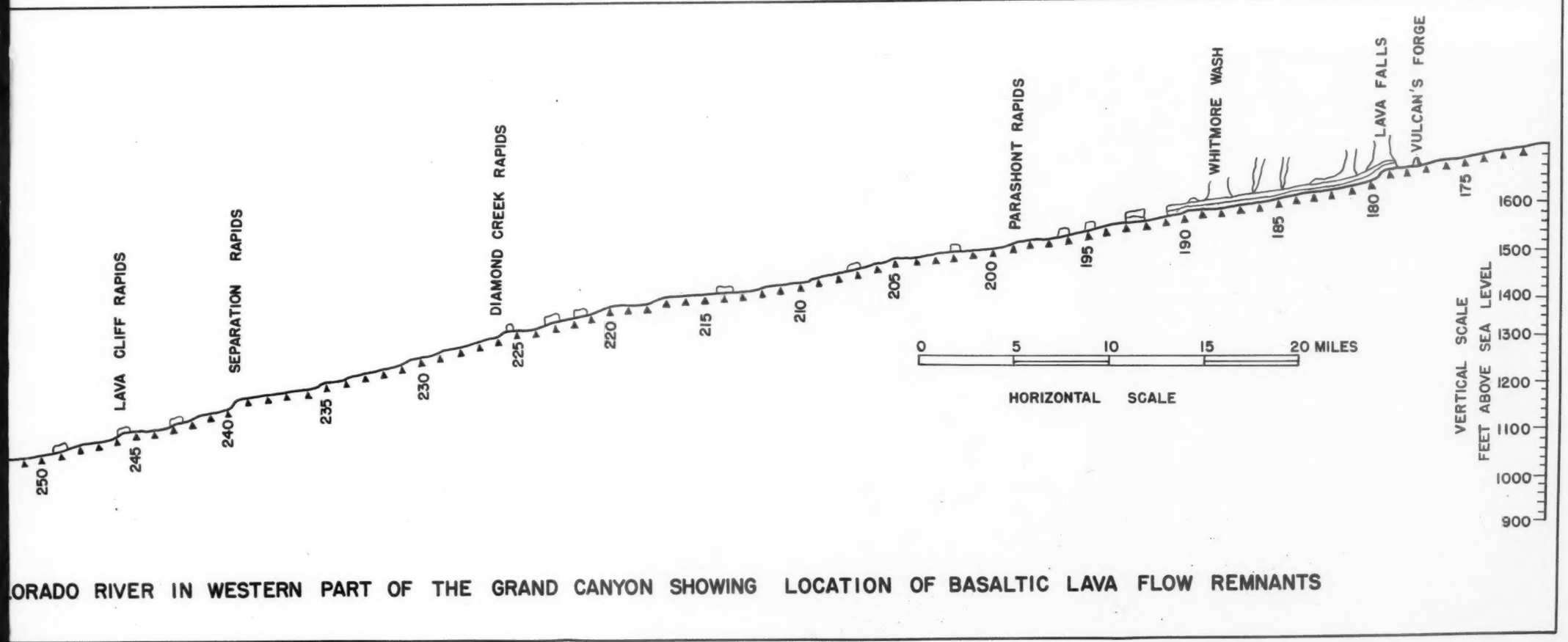

LORADO RIVER IN WESTERN PART OF THE GRAND CANYON SHOWING LOCATION OF BASALTIC LAVA FLOW REMNANTS

PLAN AND PROFILE OF THE WESTERN PART OF THE GRAND CANYON

Showing location of volcanic features

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GEOLOGIC CROSS-SECTION OF THE CORDILLERA DE LOS ANDES AT ABOUT PARALLEL 33° L.S. (ARGENTINA-CHILE)

By Félix Gonzáles Bonorino

CONTENTS

	Page
Abstract	17
Introduction	17
Paleozoic	19
Triassic	19
Middle Jurassic to Lower Cretaceous marine beds	20
Porphyritic formation	21
Tertiary	21
Structure	22
Conclusions	22
References cited	25

ILLUSTRATIONS

Figure	Page
1.—Index map	18
2.—Generalized sections showing the main stages in the geologic history of the studied area	23

Plate	Facing page
1.—Structure section of the Cordillera de los Andes and the Precordillera between Los Andes (Chile) and Villavicencio (Argentina)	24

Abstract

A profile across the Cordillera de los Andes between Villavicencio, Mendoza, Argentina, and Los Andes, Chile, shows the following stratigraphic units: (1) A basement complex composed of strongly folded beds of graywackes and slaty shales, Palaeozoic in age, slightly metamorphosed by stocks of granodiorite not younger than Middle Triassic; (2) A complex of strongly consolidated pyroclastics, intruded by large laccolithic bodies of quartz-porphyry and dikes and sills of andesites, diabase, etc. Age: Late (?) Triassic; (3) A marine series, ranging from Upper Liassic to Lower Cretaceous, formed by limestones, sandstones, anhydrite (gypsum), tuffites, shales, and conglomerates; sills and dikes of dacite, and andesite porphyries and diabase are common; (4) A thick complex of pyroclastic rocks of andesitic composition, largely breccias, traversed by dikes and sills of andesitic or dioritic rocks; partly intercalated in the upper part of the marine series. Age: Neocomian to Late Cretaceous. (5) Conglomerates and sandstones of Tertiary age. The units 2, 3, and 4 are roughly conformable with each other, and rest unconformably on 1. Unit 5 is unconformable with respect to all. The structure is characterized by a large flexure, the monoclinal part of it being parallel to and broadly coincident with the International boundary. A thrust fault transects the monocline near the base, whereby the Mesozoic beds rest partly on Tertiary conglomerates; another thrust fault marks the limit of the Cordillera to the east. The sequence of events is as follows: (1) deposition of the sedimentary-pyroclastic Mesozoic complex over an irregular peneplain of Palaeozoic rocks; (2) flexuring in the Early (?) Tertiary; (3) erosion and deposition of conglomerates and sandstones; (4) thrusting (Late Tertiary); (5) uplifting and erosion. The loftiest peak of the Cordillera, the Aconcagua (ca. 23,300 ft.), located about 20 kilometers north of our profile along the boundary, is made up of breccias belonging to unit 4 (Porphyritic formation) and structurally forms part of the huge cuesta that carries the frontier line.

Introduction

Early in 1946 the writer was commissioned by the Dirección de Minas y Geología of the Argentine government to study the geologic profile of the Cordillera de los Andes between Puente del Inca, Province of Mendoza, and the Chile-Argentine boundary, at about parallel 33° L.S. This work was intended to provide a basis for the project of a new tunnel, about 23 kilometers long, for the Transandean Railroad, between Puente del Inca and Juncal, Chile (Fig. 1).

About 25 days were devoted to the study of the stratigraphic section and the structure; the same year the observations were extended east of Puente del Inca to Uspallata and Villavicencio, aiming to obtain an overall picture of the structure of this part of the Cordillera.

The Chilean part of the profile here presented was taken from data kindly furnished by H. Flores Williams, of Chile, for between the Inter-

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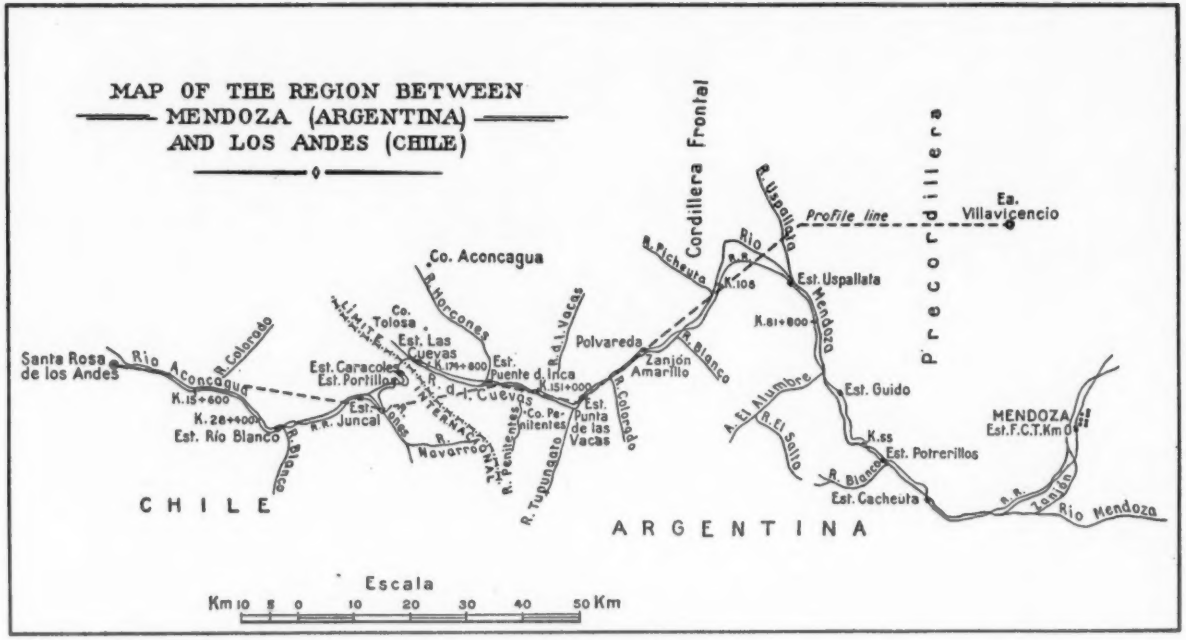


FIGURE 1.—INDEX MAP
Showing the location of the cross section

national boundary and Juncal, and from personal observations for between the latter point and Los Andes.

Charles Darwin (1846, PRV) shows essentially the same section in his "Geological observations." W. Schiller (1912) gave an account of the region between the Aconcagua and Rio Las Cuevas. Nothing but scattered references have been published since.

The profile as presented in this paper includes, for completeness, the structural unit east of the Uspallata valley, which is known by the Argentine geologists as the Precordillera.

The present study was undertaken largely on the initiative of L. R. Lambert, head of the geologic branch of the former Dirección de Minas y Geología now Dirección General de Industria Minera. He visited the writer in the field and contributed many ideas. J. J. Rossi helped in the measurement of a part of the section.

Paleozoic

The oldest member of the stratigraphic sequence shown by the profile is formed by beds of graywackes and silty shales in the Precordillera (Villavicencio) and hornfelsic graywackes and shales in the Cordillera (Zanjón Amarillo). The beds present very steep dips, and also small-scale folding, specially in the eastern area (Harrington, 1941). The thickness of the sedimentary section appears to be considerable and thus consistent with deposits of geosynclinal type.

The rocks of the Villavicencio area have been considered Devonian by Stappenbeck (1910) on very indirect evidence; very likely they belong to a geosynclinal series ranging from upper Cambrian through Devonian, and unconformably overlain by shallow marine-glacial Carboniferous sediments found elsewhere in the Precordillera. The graywackes and shales that form the core of the so-called Cordillera Frontal (Pl. 1) belong probably to the same series.

Igneous Rocks: Between Zanjón Amarillo and Quebrada de los Penitentes, the Paleozoic sediments appear intruded by a stock of medium-grained, gray biotite diorite to granodiorite, grading locally into pink granite. This body is responsible for the pronounced hornfelsic character of graywackes and shales in this area; it does not cut the overlying Triassic beds, but is separated by an erosion surface.

Triassic

Overlying the Paleozoic with strong angular unconformity there is a thick complex of pyroclastic and hypabyssal rocks, generally assumed to be Triassic. In the Precordillera (Paramillo de Uspallata), this formation consists of consolidated, light-colored tuffs, moderately well stratified, with numerous diabase sills. Thin-bedded horizons of subaqueous deposition are also present, sometimes containing fossil relics of plants and fish scales. Standing trees, first described by Darwin, are found in these beds.

The basal portion of the Triassic in the eastern slope of the Precordillera is formed by a complex of volcanic breccias, tuffs, and thick dikes of andesitic rocks, which pinches out very rapidly (Harrington and De Benedetti *in* Harrington, 1941).

The Triassic rocks, a main constituent of the Cordillera Frontal, west of Uspallata valley, are poorly bedded, indurated tuffs and breccias of different colors. The pyroclastics are intruded by various types of hypabyssal rocks, the main one being a pink quartz-porphyry that forms a large laccolithic body between Picheuta and Polvareda. Similar although smaller bodies are common elsewhere in this region. The laccolith includes in its mass isolated pyroclastic beds, showing its composite origin.

Other intrusives are albitized diabase sills and irregular bodies of andesites.

Probably much of the fine-grained pyroclastics shown in this part of the section are welded tuffs, as at the lower part of the Quebrada de los Penitentes.

The age of the pyroclastics and its intrusives west of Uspallata is taken as Triassic by lithologic analogy with the tuffaceous rocks of Paramillo de Uspallata and near-by region, where fossil plants of that period have been found.

Middle Jurassic to Lower Cretaceous Marine Beds

Between Quebrada de los Penitentes and the International boundary the main constituents of the Cordillera are sandstones, limestones, limy shales or siltstones, anhydrite (gypsum), etc., locally fossiliferous, intruded by sills and dikes of dacite and andesite porphyry and diabase. The marine beds are separated from the Triassic pyroclastics by at least a disconformity and perhaps a slight unconformity. The total thickness of this complex is nearly 5,000 meters, about 10 per cent of it corresponding to sills of various kinds.

The stratigraphic sequence, from top to bottom, is as follows:

	meters
Limestone, yellowish gray, well bedded, hard; alternating thin beds of gypsum	400
Sandy tuff, greenish gray, well bedded, partly conglomeratic	500
Limestone, hard, yellow, well bedded; locally very rich in bivalve (*Ostreidae*) shells	50
Conglomeratic sandstone, andesitic pebbles, grading into sandstones, greenish gray, tuffaceous, well bedded	150
Melaphyre sill, with conspicuous labradorite phenocrysts in dark groundmass; some copper mineralization	100
Volcanic breccia, andesitic, hard, greenish; grading downward into sandy tuff, well bedded	800
Limestone, yellowish, dark gray on fresh fracture; dark (bituminous) in the lower part. Scanty rests of molluscs, badly preserved. Sill of hornblende-andesite porphyry, 5 meters	220
Gypsum, alternating with thin limestone beds, gray, soft, sandy, and with calcareous sandstone, brownish, thin-bedded	80
Limestone, hard, dark brown, well bedded	10
Gypsum, thin-bedded, alternating with limestone, dark gray, which predominates at the base	55
Limestone, hard, dark greenish gray, thick-bedded; good marker	50
Sill of melaphyre (labradorite porphyry)	20
Limestone, dark brown, rather soft, thin-bedded	20
Sill of melaphyre, same as above	25
Sandstone, dark brown, tuffaceous; fine- to medium-grained, well bedded, variegated; wispy fragments of dark-brown shale at many levels	140
Two sills of altered diabase, fine-grained, separated by a thin bed of sandstone	60
Sandstone, tuffaceous, dark brown to brownish gray, fine-grained, hard; wispy fragments of dark-brown shale at certain levels	170
Limestone, gray, brecciated; druses parallel to flow structure	35
Anhydrite, bluish gray, banded, plastically microfolded; largely transformed into gypsum	120
Limestone, dark brown, well bedded; partly conglomeratic; alternating thin beds of gypsum	80
Limestone, superficially yellowish gray, internally creamy or dark gray; hard, well bedded. Lower part light greenish gray, with intercalated beds of coarse to conglomeratic calcareous sandstone	400
Sandstone, dark brown, fine-grained; good stratification, partly cross-bedded	300
Gypsum, whitish gray; numerous remnants of anhydrite	200
Sandstone, dark brown to yellowish brown, fine-grained, hard, laminated; very fine-grained, calcareous, fragmentose, toward the base. Alternating thin beds of yellowish limestone in the middle of the unit	100
Limy siltstone, shaly, dark brown, interbedded with limestone, yellowish, hard, laminated	100
Dacite sill, light gray, fresh; thickness variable, average about	30
Limy siltstone to fine-grained sandstone, dark brown, hard, irregularly bedded, laminated; some alternating thin gypsum beds	160
Sandstone, dark brown, medium-grained, massive	6
Sandstone, light gray, massive, grading downward into conglomeratic ss., yellowish gray, partly brownish, hard, well stratified, and conglomerate, with quartz, quartz-porphyry and andesite pebbles. Diabase sill intercalated	200
Sill of andesite porphyry, light gray, fresh, very conspicuous; thickness variable	70
Sandstone, medium-grained, dark brown, hard, laminated, often cross-bedded; locally conglomeratic; conglomerate, dark brown, subangular pebbles, aver. 5 cm., up to 20 cm., mostly porphyrites; light yellowish bands due to decoloration of the matrix. Lower part dark brown, conglomeratic ss. like top	180
Sills of diabase above and andesite below, inclosing limestone, creamy gray to light brown, with abundant relics of bivalves	50
Limestone, yellowish gray, well bedded; grades into conglomeratic limestone, dark brown	100

	meters
Conglomerate, dark to light gray, andesitic pebbles	25
Limestone, yellowish gray, partly brecciated, poorly bedded, with moulds of *Gryphaea*	170
Low-angle thrust plane	
Conglomerate, dark, andesite and limestone pebbles (Tertiary)	0 to several hundred
Unconformity	
Limestone, light gray, hard, well bedded	15
Disconformity	
Tuff and brecciose tuff, hard, greenish to whitish gray, largely ignimbritic (Triassic)	

The gypsum beds change in thickness very rapidly and taper out within relatively short distance. The thin beds alternating with the limestone of the uppermost part of the section, for instance, merge laterally to the south into a thick gypsum unit. Some of this irregularity in thickness may be original, but it is believed that most of it is due to boudinage.

In the uppermost limestone beds H. Flores Williams (1946) found relics of a mollusc, identified by J. Tavera, of Chile, as *Pecten* (*Chlamys*) *robinaldinus* D'Orbigny, which indicates Lower Cretaceous. *Gryphaea calceola* Quenst., common in the basal limestones, points to a Middle Jurassic age (Schiller, 1912).

Porphyritic Formation

This formation makes up the rest of the section west of the International boundary. It consists of strongly consolidated volcanic breccias, tuffs, and conglomerates intruded by a variety of dikes, sills, and stocks. The bulk of the complex is formed by poorly bedded, compact volcanic breccias, gray to greenish or purplish. By decrease in the number and size of the coarse elements they grade into crystal tuffs with somewhat better stratification.

The predominating magmatic type is andesitic (porphyritic); small phenocrysts of plagioclase, largely altered, are characteristic of the fragments of eruptive rocks.

Relatively well stratified conglomerates of porphyritic pebbles are found at a stratigraphic level not far from the base; they can be seen along the road between Portillo and the Chilean entrance to the International tunnel.

The pyroclastic rocks are abundantly penetrated by mesosilicic intrusives of various types[1]. A common variety is an andesite porphyry containing big hornblende phenocrysts; it is specially abundant near and at the base of the formation as dikes and irregular bodies which give an uneven appearance to the contact.

Dikes of a light-colored andesite porphyry are common throughout the complex. Some of them are spectacularly shown on the cirque walls of Cerro Tolosa and near-by peaks; they are also known from the highest part of Cerro Aconcagua, and numerous blocks of the same andesite porphyry are seen in the moraine of Rio Horcones, near Puente del Inca.

In the western part of the profile, bodies of a coarser, more even-grained rock are found, like the pyroxene diorite or microdiorite at Salto del Soldado, between Juncal and Los Andes.

The thick unit of volcanic breccia interbedded in the upper part of the marine sequence is essentially the same lithologically as the Porphyritic formation. To the north, between Cerro Tolosa and Cerro Aconcagua, the coarse pyroclastics are gradually replaced along the strike by dark-brown sandy tuff and tuffaceous sandstone.

No evidence of the age of the Porphyritic formation is available other than it closely followed, and partially accompanied, the deposition of the Neocomian beds.

Tertiary

In the Uspallata valley are preserved a series of sandstone beds, often conglomeratic, cross-bedded, gray. This formation, commonly referred to as Estratos Calchaquíes, is usually considered as middle or late Tertiary, although no direct evidence in this respect is available.

Another area of Tertiary sediments is located between Cerro Penitentes and Cerro Santa Maria. The upper part of these two peaks is formed by remnants of horizontal beds of conglomerate, lying unconformably on Mesozoic rocks. The conglomerates, which at Cerro Santa Maria display a great thickness, are typical

[1] At least some of the igneous rocks intruding the Mesozoic formations are very young (late Tertiary?). The dacitic rock found as a sill in the marine sequence can be seen cutting the Tertiary conglomerates in the valley of Santa María, between Cerro Santa María and Puente del Inca.

fanglomerates consisting of boulder and pebbles of andesites and limestone.

Structure

Combining lithologic characteristics and structure, the profile can be divided in four sections: (1) From Villavicencio to Uspallata; (2) from Uspallata to Quebrada de los Penitentes; (3) from Quebrada de los Penitentes to the International boundary; (4) from the latter to Los Andes, Chile.

Section 1 is usually referred to geographically and geologically as Precordillera; sections 2, 3, and 4 have been called Cordillera Frontal and Cordillera Principal, respectively, by Groeber (1918).

Section 1: An important fault borders the Precordillera to the east, causing the huge block of Paleozoic basement to rise with tilting to the west. Smaller blocks are found near Uspallata, and still others are probably hidden under the anticlines in the Triassic beds. The latter form on the whole a monocline that reveals the tilting of the main block. North and south of here the secondary faulting becomes more important and the Precordillera appears broken down in several subsidiary units, very conspicuous topographically.

The regional tilting of the basement is also reflected in the position of the Tertiary beds of the Uspallata valley, which dip about 25° to the west.

Section 2: Between the Precordillera and the Cordillera Frontal, a wide longitudinal depression, known here as Uspallata valley, extends from about parallel 28° to just south of Uspallata. The Cordillera Frontal forms the western side of this depression, separated from it by a large fault or system of faults. At Uspallata the main faulting has the character of overthrusting; the Tertiary beds are overridden by Triassic pyroclastics along a plane dipping about 45° to the west. Against the thrust plane, the former appear bent and overturned by dragging. This fault dies out near Uspallata, and reappears again nearly 35 kilometers south of our profile line, running along the base of Cerro del Plata.

Between the overthrust plane and Quebrada de los Penitentes the structure can be described as a very broad arch of Triassic pyroclastics changing into a flat syncline to the east. The core of the arch is occupied by the steep beds and the intrusives of the Paleozoic basement. Several high-angle faults somewhat complicate the structure; only the most important ones have been represented in the profile.

Section 3: The strata of the Mesozoic marine formation form a monoclinal structure, dipping an average of some 30 degrees to the WSW. The structure looks comparatively simple here, but as we go either north or south some complications arise, due specially to the disturbing influence of the gypsum beds. Tectonic repetition at the stratigraphic level of the main gypsum units is seen, for instance, on the north side of the valley of Río Las Cuevas, just west of Río Horcones. A few miles south of our profile line several flexures modify the monoclinal structure to a certain extent.

A main structural feature in this part of the profile is a thrust plane that crosses Río Las Cuevas at Puente del Inca, near the base of the sedimentary marine formation. This thrust was first mentioned by Schiller (1912). At Cerro Penitentes it is clearly seen how the Callovian limestones overlie the Tertiary conglomerates. The thrust plane cuts the beds at an acute angle along the strike, so that where it separates the conglomerates of Cerro Santa María from the marine beds of Cerro Almacenes, less than 15 kilometers north of Cerro Penitentes, the plane is about 2,000 meters stratigraphically higher.

Section 4: The Porphyritic formation participates also in the monoclinal structure, up to about Río Juncal; there the thick beds of volcanic breccia level off and maintain a roughly horizontal position to the end of the profile. This part of the section has been only very cursorily examined by the writer; it seems safe to assert, however, that no major faulting has taken place here.

Conclusions

The geologic composition of the Cordillera do los Andes along the present profile can be summarized as a pile of marine beds intercalated between two thick pyroclastic complexes, all Mesozoic, the whole resting unconformably on a Paleozoic basement of semi-metamorphic

pelites and graywackes and mesosilicic intrusives and being overlain unconformably by scattered remnants of Tertiary continental deposits.

Structurally it is characterized by a large-scale flexure, the monoclinal part of it being represented by section 3 and adjacent part of section 4. The uplifted part of the flexure, east of the monocline, shows a broad arch in the sedimentary pile, mostly eroded; this arching permits the exposure of the Paleozoic basement. Two thrust faults complete the tectonic picture; one, more like a reverse fault, is responsible for the uplifting of the Cordillera Frontal; the other, a low-angle overthrust, cuts the monocline of marine beds.

The geologic history of this part of the Cordillera can be sketched as follows (Fig. 2):

1. Folding, intrusion, uplift, and erosion of a middle-lower Paleozoic geosynclinal sedimentary pile, probably during late Devonian time. Formation of an irregular peneplain.

2. Eruption of large quantities of pyroclastic material, acidic in composition, followed by intrusion of hypabyssal bodies of various types and compositions, predominantly laccoliths of rhyolite porphyry. Triassic.

3. After a short period of erosion, the area begins to subside; the sea invades from the west. Limestones, sandstones, siltstones, calcareous shales, conglomerates, anhydrite, etc., about 4,500 meters thick, are deposited. Near the end of this stage there is an eruption of coarse pyroclastics, the forerunner of much more extensive and lasting volcanic processes to come.

During this period the sedimentation takes place in a shallow sea, the coast line being not very far east from Cerro Penitentes. The most spectacular proof of the shallowness of this marginal sea are the gypsum (anhydrite) beds distributed throughout the section. Eruptions of fine-grained pyroclastic material often accompanied the marine sedimentation.

4. The eruption of pyroclastics of andesitic composition extends over a wide area in the early Cretaceous, filling up the sedimentary basin and accumulating several thousand meters of breccias, tuffs, and conglomerates. Intrusion of some of the sills, dikes and irregular bodies of basic and mesosilicic rocks might conceivably have taken place then; the majority, however, were probably intruded later (late Tertiary?).

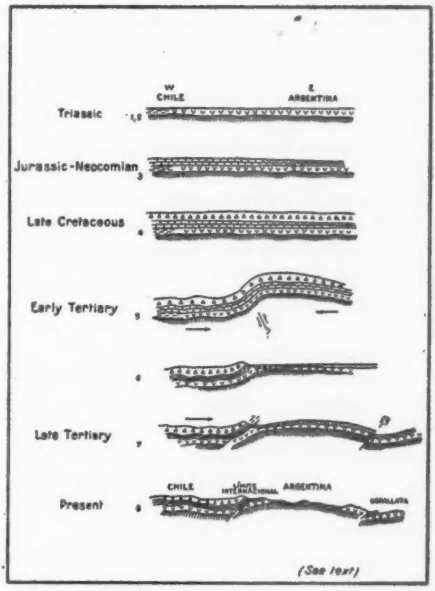


FIGURE 2.—GENERALIZED SECTIONS SHOWING THE MAIN STAGES IN THE GEOLOGIC HISTORY OF THE STUDIED AREA

The structure in the Paleozoic basement is not shown. Not at scale.

5. A strong tectonic movement, compressive in character, originates a huge flexure in the Mesozoic sedimentary pile, probably by thrusting the basement along a reverse fault. Date of this movement is uncertain, probably late Cretaceous or early Tertiary.

6. Erosion sets in immediately upon the whole region, but specially the upper (eastern) part of the flexure. Once the Porphyritic formation was removed, the erosion has no difficulties in stripping the eastern slope off the marine sediments.

The continental divide is thus established not far from its present position; piedmont sediments (conglomerates and sandstones) are deposited on both sides, the eastern ones being probably the thicker, and the only ones to be preserved. Apparently a longitudinal depression is formed along the belt of marine sedi-

ments, probably drained eastward by antecedent streams superimposed on what is now the Cordillera Frontal; in fact, the base of the Tertiary conglomerates of Cerro Santa María is well below the average altitude of the Cordillera Frontal.

7. A renewal of the compressive movements in the late Tertiary results in thrusting along a plane within the pile of marine sediments, the weakest point in the flexure. High-angle thrusting occurs also much farther east, whereby the scarp of the Cordillera Frontal and longitudinal valley of Uspallata come into being. This tectonic movement is also responsible for the uplift of the Precordillera. It is possibly at this stage that many of the hypabyssal rocks are intruded in the Mesozoic formations.

8. Regional isostatic uplift and erosion.

This sequence of events refers specifically to the Cordillera at about parallel 33°L.S., but the writer thinks that it can also be applied, in general, to the Andes of San Juan and Mendoza provinces. To the south the axis of the flexure enters well into Argentine territory, and it is still recognizable in southern Mendoza.

Although the Chilean-Argentine cordillera has often been referred to as the site of a Mesozoic geosyncline, it seems evident that, as far as the present profile is concerned, the sedimentary basin here located had the characteristics of a marginal sea rather than those of a geosyncline. Its 4,500 meters of neritic sediments can hardly be compared, in composition as well as in thickness, with the great piles of pelitic and semipelitic beds typical of the classical geosynclinal areas. As the marine deposits seem to thicken to the west (Chile), it might be that a real geosynclinal basin was located in that direction, although no evidence of it is known to the writer.

In contrast with this marginal character of the Cordilleran sea, we find in the Patagonian Andes, particularly south of parallel 42°L.S., Mesozoic sediments that indicate a rapidly sinking basin and a typical geosynclinal evolution (Petersen and González Bonorino, 1947).

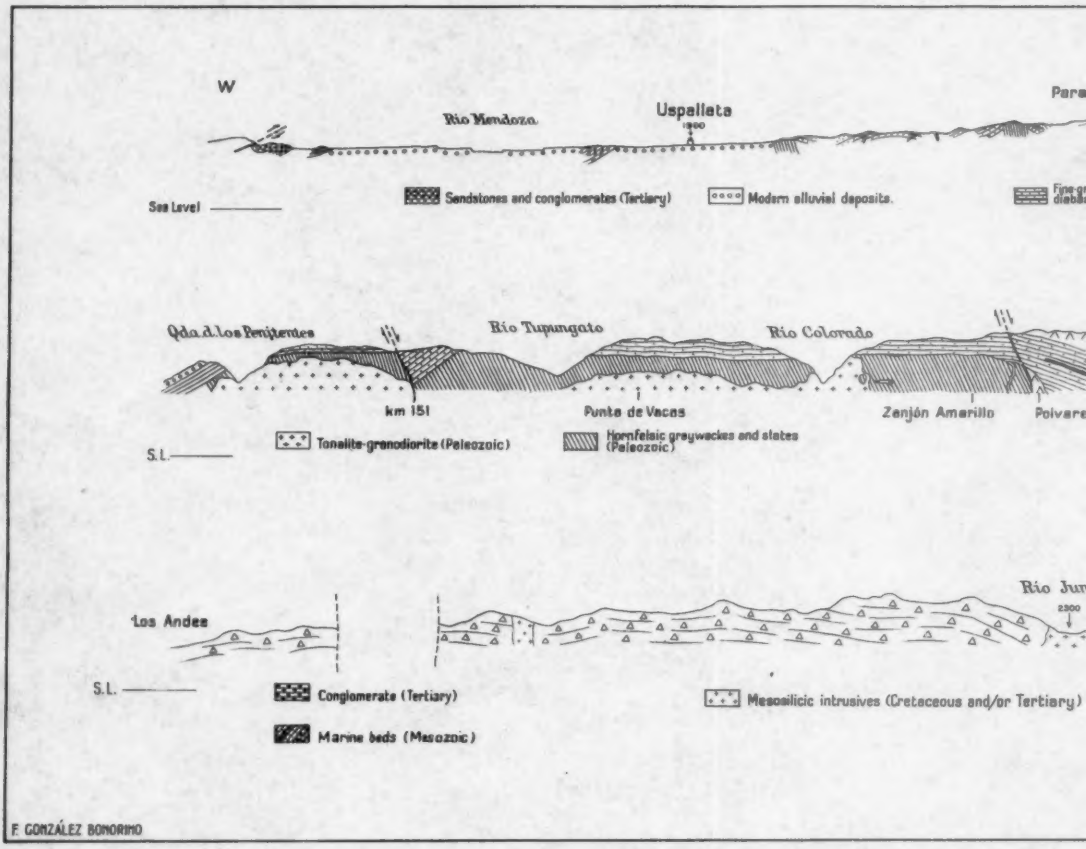
Composition and Origin of Cerro Aconcagua: The geologic nature of the Aconcagua, the highest peak (7,015 meters) in the Western Hemisphere, has of old attracted interest among the geologists of this part of America. It was at first regarded as a very much dissected volcano, of which no trace of the crater had been preserved. Its pre-eminence over the near-by peaks and its being made up largely of volcanic rocks were the foundations of this idea. Schiller (1907; 1912) opposed that interpretation with strong evidence, considering the volcanic rocks older and giving credit to the thrusting (specifically to the thrust fault of Cerro Penitentes) for the present height of the Aconcagua. Recently Fossa-Mancini (1947) has tried, on the basis of a critical review of the literature, to revive the former idea, assigning a Tertiary age to the hypothetical volcano.

The studies made in connection with the profile here presented, plus a reconnaissance trip along Río Horcones up to the base of the Aconcagua (Plaza de Mulas), have enabled the present writer to confirm Schiller's objections and obtain a reasonably generalized picture of the formation of the lofty peak.

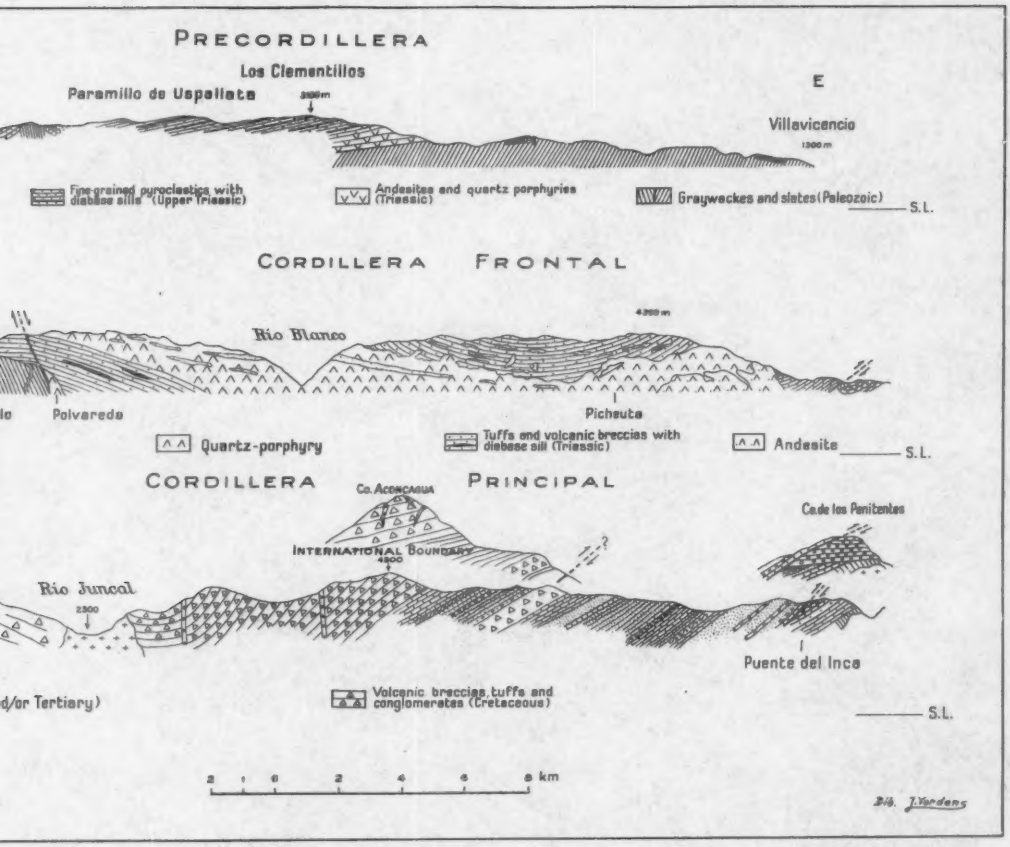
The Aconcagua is situated in the northward continuation of the huge cuesta that carries the Chile-Argentine boundary line (Fig. 1; Pl. 1). Following the valley of Río Horcones upstream, one can recognize the sequence of Mesozoic marine beds up to the Porphyritic formation, which forms the upper part of the peak as well as many of the neighbouring summits (Cerro Cuerno, etc.). The unit of volcanic breccias intercalated in the upper half of the marine sequence shown in Figure 3 forms, with a somewhat different lithologic character (see above), a prominence on the southeastern slope seen from the distance as a dark step below the steep cirque wall. This unit, incidentally, is also responsible for Cerro Tolosa.

The strongly consolidated volcanic breccias brought down by glaciers from the upper levels of the Aconcagua are identical to those forming the base and also to the ones shown in our profile as Porphyritic formation. It can be asserted that no pyroclastic rocks are found in the whole area that can not be easily correlated, structurally and lithologically, with those of the Porphyritic formation.

The reason for the extreme altitude of the Aconcagua is not easy to establish without a more detailed study; it likely lies in a number of circumstances rather than in a single cause. It must be remembered that summits of around



STRUCTURE SECTION OF THE CORDILLERA DE LOS A
BETWEEN LOS ANDES (CHILE) AND VILLAV



A DE LOS ANDES AND THE PRECORDILLERA
ND VILLAVICENCIO (ARGENTINA)

6,000 meters are common along the Andes at this latitude. The effect of the thrust fault in that connection is not known, but if we consider its behavior between Cerro Penitentes and Cerro Almacenes it seems almost certain that it will pass at the base of the eastern slope of the Aconcagua. Consequently, its participation in bringing up this part of the Porphyritic formation higher than the rest might have been effective.

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CORRELATION OF THE CRETACEOUS-TERTIARY TRANSITION IN SASKATCHEWAN AND ALBERTA

By Loris S. Russell

CONTENTS

	Page
Abstract	27
Introduction	27
Scope of paper	27
Recent work	28
Summary of stratigraphy	28
Distribution of sections	28
Southwestern Saskatchewan Section	30
Southeastern Alberta Section	31
Southwestern Alberta Section	32
Central Alberta Plains Section	33
Central Alberta Foothills Section	34
Revision of correlation	35
Southwestern Saskatchewan Section to Southeastern Alberta Section	35
Southeastern Alberta Section to Southwestern Alberta Section	36
Southeastern Alberta Section to Central Alberta Plains Section	37
Southwestern Alberta Section to Central Alberta Plains Section	38
Central Alberta Plains Section to Central Alberta Foothills Sections	39
Summary	40
Conclusions established	40
Conclusions requiring further confirmation	40
Conclusions disproved	41
References cited	41

ILLUSTRATIONS

Figure	Page
1.—Index map, Saskatchewan and Alberta	29
2.—Correlation chart, southwestern Saskatchewan to southwestern Alberta	36
3.—Correlation chart, southeastern Alberta to central Alberta plains	37
4.—Correlation chart, central Alberta plains to central Alberta foothills	39

Abstract

The author's earlier conclusions on Cretaceous-Tertiary correlations on the Canadian great plains are reviewed in the light of recent studies by various workers. The stratigraphical sequence in the five main sections of Cretaceous-Tertiary formations is summarized. Revised correlations are given between the sections in southwestern Saskatchewan and southeastern Alberta, southeastern and southwestern Alberta, southeastern Alberta and the central Alberta plains, southwestern Alberta and the central Alberta plains, the central Alberta plains and the central Alberta foothills. Most of the previous conclusions have been supported or confirmed by later work, but others are still subject to revision. The most important change required in the writer's earlier interpretation is the extension of the time range of the Edmonton formation to include a portion equivalent to part of the Lance stage.

Introduction

Scope of paper

Seventeen years ago the writer published his doctorate thesis on the Cretaceous-Tertiary transition in Alberta (Russell, 1932a). This work consisted of short descriptions of the uppermost Cretaceous and basal Tertiary formations of Alberta, and an attempted correlation of them with each other and with the time stages that were becoming recognized as standard in the western United States. Many of the conclusions were tentative and have been rightly subjected to critical scrutiny by subsequent writers, including this writer. New discoveries have been made in this field and the older stratigraphical concepts have been revised. As the immediate future will probably

see increased activity in this and related phases of Great Plains stratigraphy, it seems an appropriate time to review the writer's earlier conclusions, and to make such revision as may be required to bring them into line with the newer discoveries.

Recent work

The pioneer work of Sanderson along Red Deer River, completed when the writer's investigation had only just begun, has at last been published (Allan and Sanderson, 1945). Sanderson's results, fortunately, were available to the writer during the preparation of his original paper, and necessitate no major changes in his conclusions. However, the publication of Sanderson's observations gives the full documentation for the physical evidence on the Cretaceous-Tertiary unconformity in central Alberta, as well as providing a standard section for the important Edmonton formation.

The most important work to follow the writer's original paper was the Memoir on the geology of southern Saskatchewan (Fraser *et al.*, 1935), which, although bearing the names of five authors, was primarily the work of F. H. McLearn and P. S. Warren. Here the Saskatchewan development of the Cretaceous-Tertiary transition was fully described, and the peculiar position of the main unconformity was demonstrated.

The next major discussion of the subject appeared in the Memoir on the geology of the southern Alberta plains (Russell and Landes, 1940). Correlation between the sections of southern Saskatchewan and that of southeastern Alberta was postulated, and a very late marine Cretaceous fauna was described. The gap between McLearn's area on the east and that of Russell and Landes on the west was closed by Furnival (1946) in his study of the Cypress Lake map area. More detailed east-west correlations were thus made possible. Minor contributions in the area were the discovery of a late Eocene mammalian fauna (Russell and Wickenden, 1933) and the finding of a determinable marine fauna in the type Eastend formation (Russell, 1943). During the field season of 1948, the writer studied the Late Cretaceous and Early Tertiary strata in the Cypress Hills area of Saskatchewan, and some of the data obtained during this work have been used in the final revision of this paper. For permission to do this the writer is indebted to the Department of Natural Resources and Industrial Development, Province of Saskatchewan.

Farther west the detailed structural studies of G. S. Hume in the southern foothills and of B. R. McKay and A. H. Lang farther north revived consideration of the Cretaceous-Tertiary problems along the disturbed belt. No definite conclusions had been reached, however, until the fossil plants were studied by W. A. Bell. His investigation established the much lower position of the Cretaceous-Tertiary boundary than had previously been recognized (Lang, 1947).

Numerous excellent contributions to the stratigraphy of the Cretaceous-Tertiary transition have been made recently by workers in the western United States. The paleobotanical works of Dorf (1938, 1940, 1942a, 1942b) and Brown (1939, 1943) have cleared away many of the supposed discrepancies in the evidence of the fossil plants. Relationships between the vertebrate faunas and the stratigraphy have been established by Simpson (1937) in southwestern Montana, Jepsen (1930, 1940) in northwestern Wyoming, Gazin (1941a) in central Colorado, and Gazin (1941b) and Gilmore (1946) in central Utah. Particularly noteworthy is the general discussion of the Cretaceous-Tertiary problem in Spieker's (1946) paper on central Utah.

Summary of Stratigraphy

Distribution of sections

Many of the difficulties of correlation in the plains of Saskatchewan and Alberta (Fig. 1) arise from lateral discontinuity of sections. This results from a combination of structural and topographical conditions. In the western part of Alberta, adjacent to the Rocky Mountain front, there is a broad, asymmetrical trough known as the Alberta syncline. Its axis is occupied by the Tertiary formations; to the west the Cretaceous strata appear at the surface with strong dips to the east; on the east side the westward dips of the underlying formations are almost imperceptible. In southwestern Alberta the

syncline is so narrow that there are no essential changes in the younger formations from east to west. For present purposes, therefore, we need recognize only one sequence here, the Southwestern Alberta Section. Farther north, where the Alberta syncline widens and becomes almost a monocline, marked differences occur in the Cretaceous-Tertiary transition as exposed on the east and on the west side. We distinguish here between a Central Alberta Plains Section and a Central Alberta Foothills Section. The eastern limb of the Alberta syncline passes into a broad arch, trending somewhat east of north across southern Alberta and losing its distinct nature in the east-central Alberta region. This structure, known as the Sweetgrass arch, has brought the older portions of the Montana group to levels where atmospheric erosion has exposed them. Thus the latest Cretaceous, and the Tertiary formations, have been removed from the axial portion of this arch, but have been preserved on the eastern flank, in the southeastern corner of Alberta. Here too the peculiarities of the drainage, and also perhaps the nature of the sediments, have preserved the high erosion remnants of the Cypress hills, affording us a more extensive view of the Cretaceous-Tertiary transition than would have been provided by the structural conditions alone. The sequence of this area may be designated the Southeastern Alberta Section.

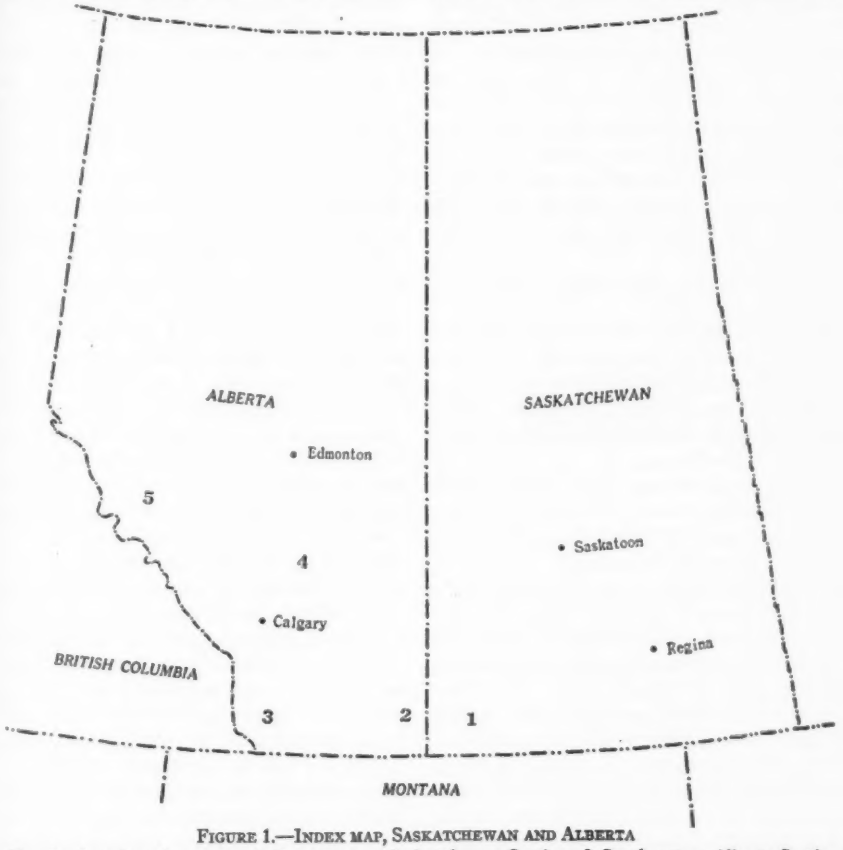

FIGURE 1.—INDEX MAP, SASKATCHEWAN AND ALBERTA

Showing location of sections: 1, Southwestern Saskatchewan Section; 2, Southeastern Alberta Section; 3, Southwestern Alberta Section; 4, Central Alberta Plains Section; 5, Central Alberta Foothills Section.

The Cypress hills extend eastward into Saskatchewan and provide many additional glimpses of the strata under discussion. Certain lateral changes can be observed, and in most cases the relationship is clear. Lack of complete exposure, however, leaves a few unsolved problems, and for this reason it is convenient to distinguish a Southwestern Saskatchewan Section. The characteristic features of this section persist eastward into south-central Saskatchewan, but are not recognizable in the southeastern corner of the province (Estevan coal field), or in the Turtle Mountain area of Manitoba.

Southwestern Saskatchewan Section

Furnival (1946) showed that the Bearpaw formation of this area, typically dark-gray marine shale, includes in its upper portion several distinct sandstone members. These members usually consist of fine-grained, clayey sandstone or siltstone, only moderately indurated, usually with an abrupt upper boundary, but passing downward into the shale by a transitional zone. Thicknesses range from 30 to 50 feet near the Alberta boundary but decrease to zero about 40 miles east, where the sandy beds are replaced entirely by shale. Furnival recognized three such members, which he named, from oldest to youngest, the Oxarart, Belanger, and Thelma. The recent work of the writer (in press) in this area has shown that the Thelma sandstone does not exist as a distinct member in Saskatchewan, that the Belanger and Oxarart sandstones are distinct and persistent members, and that there is a fourth, thick sandstone below the Oxarart, equivalent to the sandstone in Alberta erroneously called Oxarart by Furnival.

Above the highest sandstone member the Bearpaw consists of the characteristic gray shale, which merges through a few feet of sandy shale into the predominantly sandstone formation known as the Eastend. In this area the Eastend is a rather uniform formation, about 100 feet thick, consisting mainly of very fine, soft sandstone, light buff to yellowish, in part with fine clay banding. It contains beds of limonite concretions and fine layers of carbonaceous matter. Fossils are rare, but at one locality a rich pelecypod fauna occurs (Russell, 1943), apparently the youngest marine fauna of the Canadian plains, and equivalent in age to the Fox Hills fauna of North Dakota. The fine yellow sand of the Eastend passes upward, with a transitional contact, into the gray and white kaolinized sands and clays of the Whitemud formation, the most conspicuous component of this section, and one that is quarried locally for refractory clay. The thickness of the formation, as now defined, is 50–75 feet. It contains only obscure plant and pelecypod remains but appears to be the northern extension of the Colgate sandstone, commonly regarded as the upper part of the Fox Hills series. Formerly a zone of purplish-gray, bentonitic clay lying above the white beds was included in the Whitemud formation, but Furnival (1946, p. 89) has separated this as a distinct formation, the Battle. The thickness of the Battle formation is about 25 feet, except where part or all has been removed by pre-Frenchman erosion.

The rocks above the Battle were formerly grouped together as the Ravenscrag formation (McLearn in Fraser *et al.*, 1935), although it was recognized that the lower portion was of Cretaceous age as distinct from the remainder of probable Tertiary age. Furnival (1946, p. 94) separated this lower, Cretaceous part as the Frenchman formation, and gave the diagnostic features as "thick, massive or coarsely cross-bedded, medium-grained sandstones with rare coaly beds." Reference to McLearn's description (Fraser *et al.*, 1935, p. 41) of the "Lower Ravenscrag" shows that this is only one of two distinct lithological phases characteristic of this portion of the sequence. In addition to the sandy phase described by Furnival, originally called Sand B by McLearn, there is also a phase characterized by "dark- and light-coloured plastic clays and shaly silts, bentonitic in places" (McLearn in Fraser *et al.*, 1935). McLearn states that the sandy "zone" is the lower part of the "Lower Ravenscrag," and is especially well developed where there has been erosion of the Battle beds, whereas the clay "zone" represents the upper part. This may be the relationship in places, but where observed by the writer in adjacent outcrops, the two phases appeared to be either stratigraphically equivalent, or to have the sandy phase lying above the clay phase. Dinosaur bones, common

in the clay phase, have not been found in the sand, so that it is not certain that the sand is Cretaceous. An alternative interpretation, which requires further evidence for proof, is that the sand phase of Furnival's Frenchman formation represents very early Tertiary channel filling in the various Cretaceous rocks, including the dinosaur-bearing, clay phase of the Frenchman.

Whether or not the sandy beds are to be included within the Frenchman formation, this unit rests unconformably upon Battle and older sediments. Along the southeastern flank of the Cypress hills this unconformity appears as a moderately fluctuating plane of contact, and a recognizable portion of the Battle formation usually persists below. Farther southeast, down Frenchman River valley, the base of the Frenchman formation can be seen cutting down progressively through the Battle, the Whitemud and the Eastend, until finally it reaches a stratigraphical level well down in the Bearpaw formation. Thus a vertical interval of over 200 feet is transgressed in a distance of about 20 miles. As it cuts down, the Frenchman formation thickens, so that the contact with the overlying Ravenscrag formation maintains approximately its stratigraphical position. These transgressive beds are of the clay phase and, like the similar sediments farther west, contain dinosaur remains. Skulls and bones of the horned dinosaur *Triceratops* have been found here, indicating correlation with the Lance stage of the Late Cretaceous. The Frenchman formation, in the writer's view, is the Canadian equivalent of the Hell Creek beds, which also contain the Lance dinosaur fauna, and are known to rest in places unconformably upon the Colgate sandstone (Dobbin and Reeside, 1929).

The Ravenscrag formation, in the restricted sense defined by Furnival and here adopted, consists of fine sandstones, clays and shales, light-gray to buff with much interbedding and lateral change. Lignite beds, many of them workable coal seams, occur throughout the formation, and the lowest of these is taken provisionally to mark the base. In south-central Saskatchewan, a white, kaolinized member (Willowbunch), over 400 feet above the base, has been mistaken for the Whitemud formation. The total thickness of the Ravenscrag formation varies from place to place, because the top of each section is an erosion surface, but the average for southwestern Saskatchewan is about 200 feet. The fossil plants and mollusks of the Ravenscrag formation are similar to those of the Fort Union group, and the occasional bone bed contains no trace of dinosaurs. Mammalian fossils have not yet come to light in the formation, but an assignment to the Paleocene may be made with some confidence. The lower part of the formation is predominantly gray, the upper buff. The gray phase is comparable with the Ludlow and Lebo formations of eastern Montana and adjacent North Dakota; the buff phase is almost identical lithologically with the Tongue River formation.

In the Cypress Hills area, the Ravenscrag is overlain by the conglomeratic sandstones of the Cypress Hills formation, of Early Oligocene age. Farther east the upper limit is capped by the Wood Mountain gravel, of Miocene age.

Southeastern Alberta Section

This is the continuation into Alberta of the sequence surrounding the Cypress hills and shows a close resemblance to the section of southwestern Saskatchewan. However, it is also transitional to the sections farther west in Alberta. Three distinct sandstone members occur in the upper Bearpaw here. Russell (1932a, p. 132; Russell and Landes, 1940, p. 86) included them in the Eastend formation, along with the intercalated shale beds, but Furnival (1946, p. 74) very properly assigned them to the upper part of the Bearpaw. Furnival named the upper sandstone the Thelma member, and correlated the middle and lower sandstones with the Belanger and Oxarart, respectively, of southwestern Saskatchewan. The writer (in press) has found that the middle sandstone of the Southeastern Alberta Section is equivalent to the combined Oxarart and Belanger members of southwestern Saskatchewan, and that the lower sandstone, over 100 feet thick, is a distinct member, about 40 feet below the true Oxarart of Saskatchewan. For this fourth or lowest sandstone member, the writer is proposing the name Black Eagle sandstone. The

type section occurs on the Black Eagle ranch south of Thelma, Alberta, but the member extends at least as far east as Battle Creek, Saskatchewan. Well-preserved marine mollusks have been found in the Bearpaw sandstones of southeastern Alberta, particularly in the lower part of the middle, or combined Oxarart-Belanger member. This fauna was described by Landes (Russell and Landes, 1940, pt. II) as from the Eastend formation, following Russell's inclusive usage, but it is much older than the Eastend as now defined.

The Eastend formation of southeastern Alberta, in the restricted sense, consists of fine, buff-colored sandstones in the lower and upper parts, but with a middle portion of variegated sands and clays with at least one workable coal seam. Passage to the Whitemud is very transitional. The Whitemud formation, in the restricted sense used by Furnival, is only about 10 feet thick here, and is predominantly silty, with only an occasional bed of refractory clay, but, nevertheless, characteristically white. The overlying gray and brown clays, corresponding to the Battle formation, may reach a thickness of 30 feet, but they are commonly much thinner, due to post-Battle erosion. Where complete, the Battle beds contain near their top a persistent bed, about 6 inches thick, of dense, light-gray, siliceous rock, identified by Sanderson (1931) as the Kneehills tuff of the Central Alberta Plains Section.

The post-Battle unconformity may eliminate beds here as far down as the upper part of the Eastend formation. Above the unconformity the rocks are predominantly brown to buff, fine-grained, crossbedded sandstones, with indurated masses present locally. The writer (1932a) assigned these sandstones to the Ravenscrag formation, with the suggestion that they represented only the upper or buff phase of the Saskatchewan Ravenscrag, the gray phase, and the dinosaur-bearing "Lower Ravenscrag" (Frenchman) being absent. Furnival (1946, p. 104) disagreed with this and pointed out the resemblance between this supposed Ravenscrag of Alberta and the sandy beds of his Frenchman formation in Saskatchewan. This question will be discussed further under correlation, but the possibility of the sandy beds being younger than true Frenchman suggests a way in which the stratigraphical and paleontological evidences might be reconciled. Total thickness of the post-Battle sediments in the Cypress hills of Alberta is 560 feet. These sediments are overlain by about 25 feet of coarse conglomerate, evidently the western extension of the Cypress Hills formation, but more purely conglomeratic, and apparently without fossil content.

Southwestern Alberta Section

The typical development of this sequence is in the southwestern corner of the province, between Lethbridge and Waterton lakes; the exposures are found along Oldman, Waterton, Belly, St. Mary, and Milk rivers. The Bearpaw formation has no sandstone members near the top, but passes with some alternation of sandstone and shale into the massive, buff sandstone of the Blood Reserve formation. This is 50 to 100 feet thick and is the northern continuation of the Horsethief sandstone of northwestern Montana. It contains a few marine invertebrates of the Bearpaw fauna in its interbedded shales, and usually terminates above in a conspicuous bed of brackish water fossils. The overlying formation is the St. Mary River, more than 1,000 feet of light-gray calcareous sandstones and gray or greenish mudstones, with great lateral variability. There is an abundant and characteristic fauna of fresh-water mollusks, and numerous occurrences of isolated dinosaur bones. The contact with the overlying Willow Creek formation is usually sharp. The latter is composed of silts and impure clays, with few hard beds. These sediments weather readily into "badlands" forms. The most characteristic feature is the color, which is predominantly pink but includes pastel shades of maroon, purple, and blue. North of Oldman river these brilliant colors fade out. The formation is about 1,000 feet thick and is largely unfossiliferous. The upper portion, on Willow creek, contains fresh-water mollusks similar to those of the Fort Union fauna, and the entire formation was referred tentatively (Russell, 1932a) to the Paleocene. Some later workers have questioned this assignment, and Furnival (1946, p. 89) has emphasized the analogy in position of the Willow Creek and the Whitemud. The uppermost formation of the South-

western Alberta Section is the Porcupine Hills, a thick and highly lenticular series of massive calcareous sandstone, usually with clay-pellet conglomerates, and a small amount of shale. The contact with the Willow Creek formation is transitional, with much interbedding and lateral change.

Central Alberta Plains Section

The typical development of this sequence occurs east and northeast of Calgary, with good exposures on Bow and Red Deer rivers. Here the uppermost portion of the Bearpaw formation consists of brown shales, with thin beds of fine, clayey sandstone (Bassano member). The contact with the overlying Edmonton formation is distinctly marked by the appearance of light-gray, bentonitic sandstones. Oyster shells usually occur in the vicinity of the contact. The typical Edmonton sediments are pale-gray, soft sandstones which weather readily into "badlands" forms. With these may occur many irregular beds of brown and gray clay and indurated calcareous and ferruginous masses. Coal seams are numerous and persistent. The ferruginous and carbonaceous matter give the outcrops a typical color-banding of gray, rusty, and dark gray. In the upper part is a zone of white and somber clays and sands that closely resemble the sediments of the Whitemud and Battle formations and are commonly accepted as their equivalents. The topmost beds, where present, are predominantly buff sandstone, rather free of clayey material. Maximum thickness of the formation is slightly over 1,000 feet, but it is commonly less as a result of post-Edmonton erosion. The fossils include plants, brackish water and fresh-water mollusks, and a rich fauna of vertebrate remains, especially those of dinosaurs. The dinosaurian fauna, as made known by Lambe, Brown, C. M. Sternberg, and other workers, shows, for the most part, a definite pre-Lance characteristic, but very recently Sternberg (1947) has discovered that the uppermost beds of the formation, lying above the Whitemud and Battle equivalents, contain a characteristic Lance dinosaur fauna. The writer (1932a) previously assigned the Edmonton formation, on stratigraphical and paleontological evidence, to an age in part equivalent to late Pierre, in part representing the Fox Hills stage. It is now evident that the time range of the Edmonton formation overlaps that of the Lance stage.

Overlying the Edmonton formation in the central Alberta plains area is the Paskapoo formation, which almost all workers have recognized as Early Tertiary. The Paskapoo usually rests upon the Edmonton beds with an abrupt contact, marked by a basal sandstone of thick and massive character. Although not strictly conglomeratic, this sandstone contains pebbles derived in part from the underlying Edmonton beds but including fragments from igneous masses and from Paleozoic sediments. Sanderson (Allan and Sanderson, 1945) along Red Deer River first established the presence of an erosional unconformity at this contact; his surveys have indicated that as much as 450 feet of Edmonton sediments have been removed in places before the deposition of the basal Paskapoo sandstone. Occasionally, however, the characteristic basal Paskapoo sandstone is absent, and the Edmonton passes into the Paskapoo with a much less abrupt transition. Such a condition characterizes those places where the Edmonton formation was little eroded prior to the deposition of the Paskapoo sediments; in these erosion remnants of uppermost Edmonton the Lance dinosaurs occur. Above the basal sandstone, the Paskapoo, very characteristically, consists of irregular alternations of massive to thin-bedded calcareous sandstone with layers of non-laminated friable gray and brown shale. Although channelling and cross-bedding are locally conspicuous, the marked lateral change characteristic of the Paskapoo strata may not be obvious unless there is a large exposure. Few if any beds in this formation are suitable for long-distance correlation. Throughout most of its occurrence, the Paskapoo formation forms the bedrock immediately underlying the soil and drift, and its original total thickness cannot be estimated. Adjacent to the disturbed belt in western Alberta, at least 3,000 feet of Paskapoo sediments appear to be present. The fossil content of the formation is not as rich as that of the underlying Edmonton, but well-preserved fossil plants of Fort Union type frequently occur, and abundant non-marine mollusks, both fresh-

water and terrestrial, of the typical Fort Union fauna. Vertebrate remains are less common, but along with fragmentary bones of turtles, *Champsosaurus*, and crocodiles, mammalian teeth are occasionally found (Simpson, 1927; Russell, 1926, 1929, 1932b). Although occurring at various levels within the formation, those teeth that are diagnostic of exact age all indicate a correlation with the Late Paleocene, that is, the Tiffanian and Clarkforkian of the western United States. If the Paskapoo formation is exclusively of this age, then the time interval represented by the Edmonton-Paskapoo unconformity can be equated with Early and Middle Paleocene time, and possibly with part of the latest Cretaceous (Lance stage).

Central Alberta Foothills Section

Within the disturbed belt, along the western margin of central Alberta, from Bow river to north of the Athabaska, the younger strata display a more extensive development of the fresh-water facies than in any other section within the region. The youngest marine rocks of this area are of early Montanan age, corresponding to the Telegraph Creek formation of the northwestern plains States. Above these marine shales lies a great thickness of purely continental strata, possibly 13,000 feet thick in places (Allan and Rutherford, 1934, p. 36). Various names have been applied to this great non-marine series. In 1911 Malloch used the name Brazeau for the lower part. Allan and Rutherford (1923, p. 51) designated the whole as the Saunders series. Later these same authors (1934, p. 34) applied the more inclusive term Foothill series to these and similar sediments. The general correlation of the series has long been recognized by all workers. It covers an age range extending from early Montanan time to the close of the Cretaceous, or later. It is therefore the equivalent of several highly varied series of marine and non-marine strata lying farther to the east and south. Alternating sandstone and shale predominate. The sandstones are of medium to coarse grain, and occasionally conglomeratic. As in the case of the Paskapoo sandstones, they lens out laterally within short distances. The shales are also of the fresh-water type, poorly laminated, rather friable, usually with an appreciable sand content, the kind of argillaceous sediment frequently designated mudstone. Coal seams are common, especially in the middle portion of the series, and in many cases reach commercial thicknesses. The fossil content of the series is not rich, but includes plants, invertebrates, and vertebrates. The plant fossils are more widely distributed throughout the series than are the other organic remains, and include floras of both Cretaceous and Tertiary age. The non-marine molluscan faunas are usually confined to the upper part of the series, although in the more southerly portion of the area a few rich beds of fresh-water shells occur in the lower part. Along with abundant aquatic forms, such as pond snails and river clams, there is a conspicuous element of land snails. Except for the fossils in the lower beds, these non-marine mollusks are closely comparable with the Fort Union molluscan fauna, and are taken to indicate a Paleocene age. Of the vertebrate fossils, bones of dinosaurs are found occasionally in the lower part of the series. Most commonly, however, the vertebrate remains consist of fragmentary fish skeletons, or the bones of reptiles other than dinosaurs. A few mammalian teeth have been found in the middle and upper portions of the series, and all of the significant ones indicate a Paleocene age.

Although the general correlation of the Late Cretaceous and Early Tertiary sediments of the Central Alberta Foothills Section is well established, there is still considerable disagreement as to the detailed equation with the formations to the east. The evidence of the fossils is strongly in support of the view that the thick upper portion of the series is of Early Tertiary age. Thus Warren (*in* Rutherford, 1926, p. 12, 13) recognized the Tertiary aspect of the plants and invertebrates in the upper part of the series. Russell (1932a, p. 146, 147), on the basis of the non-marine molluscan faunas, extended the range of the Tertiary portion to include the uppermost part of the coal-bearing division, and (1932b, 1948) has also recorded the presence of Paleocene mammals at various horizons within the series. The paleobotanical studies of Bell (*in* Lang, 1947, p. 32) indicate that Tertiary correlation must be applied even farther down in the series, to include most of the commercial coal seams and their associated strata.

Revision of Correlation

Southwestern Saskatchewan Section to Southeastern Alberta Section

In earlier studies of the section in the Cypress hills of southeastern Alberta, Russell (1932a, p. 132, 133) included within the Eastend formation a thick series of sandstones and shales, in part marine, which he stated to be equivalent to upper Bearpaw beds of Saskatchewan, and therefore below the Eastend formation of the type area. Furnival (1946, p. 76) actually demonstrated this relationship, showing that the sandstone members in the upper part of the Bearpaw formation in Saskatchewan merged into shale eastward, whereas to the west they were the sandstones included by Russell in the Eastend formation. Accordingly, Furnival recognized these sandstones, both in southeastern Alberta and in adjacent Saskatchewan, as members of the Bearpaw formation, and named them in ascending order the Oxarart, Belanger, and Thelma sandstones. The Thelma sandstone, best developed in Alberta, was thought to extend as far east as Battle creek, Saskatchewan, and the Belanger and Oxarart sandstones to be traceable from Alberta eastward for about 36 miles into Saskatchewan.

The recent field work of the writer (in press) corroborates the general relationships of these sandstone members as described by Furnival, but shows that the details are somewhat different than postulated by that author. Thus the lower sandstone of the Southeastern Alberta Section reappears on Battle Creek, Saskatchewan, 40 feet below the true Oxarart member, with which it was correlated by Furnival. For this lowest sandstone, the writer is proposing the name Black Eagle member. The middle sandstone of the southeastern Alberta section, on the basis of thickness and position relative to other members, is the equivalent of both the Oxarart and Belanger sandstones of Saskatchewan, as well as the intervening shale beds (Fig. 2). Landes' "Eastend" fauna (Russell and Landes, 1940, pt. II) can now be recognized as occurring in beds equivalent to the Oxarart sandstone of Saskatchewan, over 200 feet below the Eastend formation as now restricted. As to the Thelma sandstone, a distinct member in southeastern Alberta, there are sandy beds along Battle creek that Furnival recognized as the eastward extension of this member, but which the writer does not regard as distinctive enough for such a correlation.

The Eastend formation, as now restricted, maintains its thickness of about 100 feet from east to west. Its rather uniform lithological character in Saskatchewan is altered in Alberta by the appearance of a middle member with shale beds and coal seams. Traces of this middle member can be found in the Eastend of the Fort Walsh district, Saskatchewan. The Whitemud formation can be restricted in Alberta, as in Saskatchewan, to the light-colored clays and sands, placing the overlying gray and brown clays in the Battle formation. In southeastern Alberta the Battle formation, and even part or all of the Whitemud formation, may have been removed locally by pre-Ravenscrag erosion. The sediments above the unconformity were referred (Russell, 1932a, p. 142) to the Ravenscrag formation and equated with the upper or buff portion of the type Ravenscrag of Saskatchewan. This correlation implies that the rocks overlying the post-Battle erosion surface are progressively younger from east to west. Some support is given to this view by the section in the vicinity of Fort Walsh, near the western edge of Saskatchewan. Here the Battle clays are overlain unconformably by gray and brown impure clays and sands, partly bentonitic, with thin coaly beds. Lithologically these strata resemble the somber lower portion of the type Ravenscrag (Middle Ravenscrag of McLearn), and appear to be without dinosaur remains. It is concluded that the post-Battle erosion surface is overlain by Cretaceous (i.e., Lance) beds of the Frenchman formation about as far west as Ravenscrag village. Between here and the Alberta boundary the overlying strata represent the lower, somber portion of the restricted Ravenscrag formation; farther west only the upper, buff portion of this formation is present. Such a correlation is supported by the presence, near Eagle butte, Alberta, of a nonmarine molluscan faunule in basal Ravenscrag beds. The fossils, originally reported by Dyer (Williams and Dyer, 1930, p. 68) and listed in revised determination by the writer (1932a, p. 142), clearly represent species of the Fort Union fauna and those found in the restricted Ravenscrag of Saskatchewan.

This question of the upper limits of the post-Battle unconformity from east to west is important to the physical history of the western plains. Unfortunately, the writer's views still require complete confirmation and are directly opposed by the correlations of Furnival, who (1946, p. 105) has identified the sands in the lower part of the Alberta Ravenscrag as the equivalent of his Frenchman formation farther east. In the descriptive section, above, it has been pointed out that this formation as delimited by Furnival shows two distinct lithological phases, one predominantly sandstone, the other predominantly bentonitic clay. Dinosaurs of the Lance fauna occur in the clay phase, and it is possible that this alone is of Cretaceous age, the sandstone being channel fillings of very early Tertiary date. In any case, these sandstones are not continuous from east to west, and the resemblance may be merely the result of depositional conditions on the post-Battle land surface. The Fort Union molluscan faunule in the basal Ravenscrag of Alberta, mentioned above, is strong evidence against the correlation of these beds with the Frenchman of Saskatchewan. It is concluded that the evidence still favors the writer's original correlation of the Alberta Ravenscrag as Paleocene exclusively.

Southeastern Alberta Section to Southwestern Alberta Section

Most students of the stratigraphy of southern Alberta now recognize that the upper limits of the marine Bearpaw formation occur progressively higher in the sequence from west to east across this region (Fig. 2). Furnival (1946, p. 63) equates the "Oxarart" (Black Eagle) sandstone of southeastern Alberta and adjacent Saskatchewan with the Blood Reserve sandstone of the Southwestern Alberta Section, a formation that immediately overlies the Bearpaw. On this basis, some 200 feet of uppermost Bearpaw strata on the southeast would be the time equivalent of entirely post-Bearpaw sediments on the west. From paleogeographical considerations, one might go further and postulate that the Blood Reserve sandstone is somewhat older than the Black Eagle, although possibly once continuous with it across the area now eroded to expose the older formations. That is to say, the Blood Reserve and Black

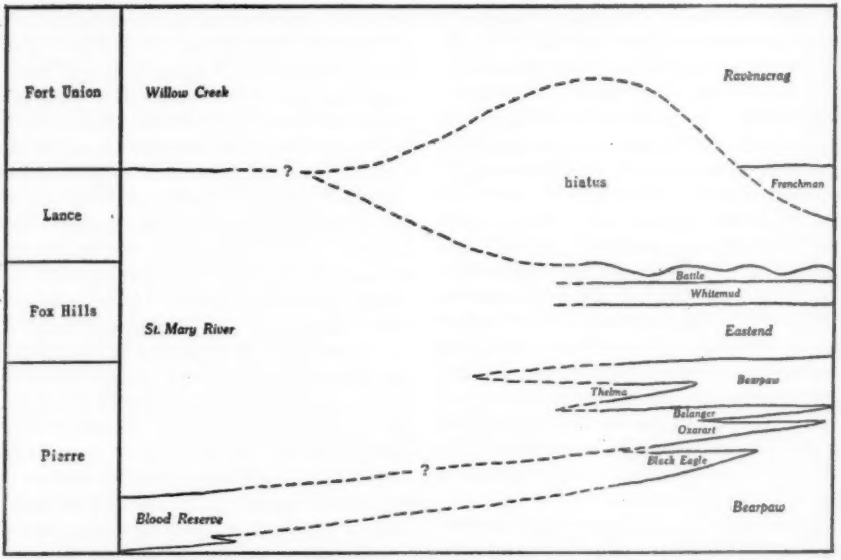

FIGURE 2.—CORRELATION CHART, SOUTHWESTERN SASKATCHEWAN TO SOUTHWESTERN ALBERTA
Column on left shows approximate equivalents in northern U. S. Plains region; not to scale.

Eagle sandstones may represent the two ends of a once continuous zone of deltaic deposits, which moved gradually eastward during the withdrawal of the Bearpaw sea. This hypothesis leads to the suggestion that the equivalents of the uppermost Bearpaw and the Whitemud and Battle formations of the southeast are to be found in part of the St. Mary River formation of southwestern Alberta. It now appears likely that Lance equivalents are also present in the St. Mary River formation. Sternberg (1947, and personal communication) has found the small ceratopsian dinosaur *Leptoceratops* to be restricted to the Lance equivalent in the Edmonton formation of central Alberta. Some years ago an incomplete skeleton of this dinosaur was collected by Barnum Brown from the St. Mary River formation of northern Montana. A Lance correlation for the uppermost St. Mary River beds would equate them with the Frenchman of Saskatchewan, and would rule out the suggestion (Furnival, 1946, p. 89) that the overlying Willow Creek formation is in part equivalent to the Whitemud. A Lance correlation for the lower part of the Willow Creek formation is possible, but still without supporting evidence. Correlation of the upper part of this formation with the Paleocene, originally based upon the presence of a Fort Union molluscan fauna, has been confirmed by the paleobotanical studies of W. A. Bell (personal communication).

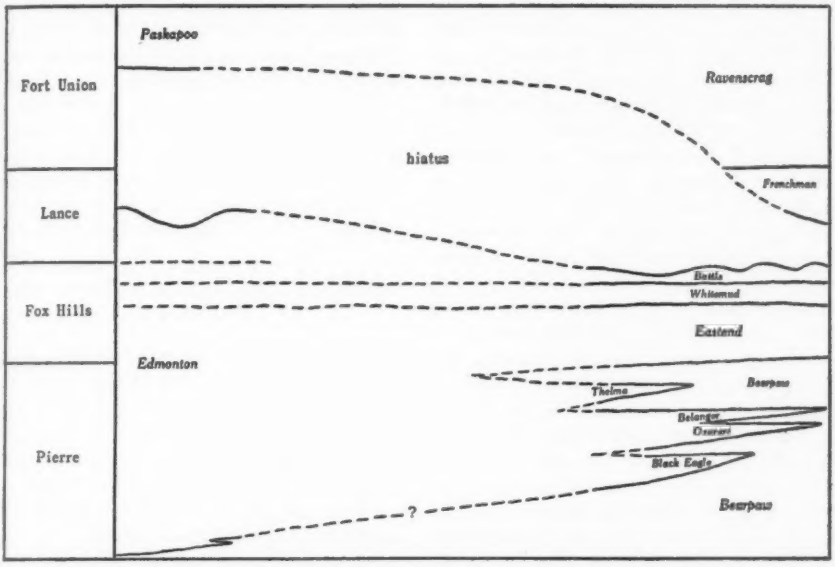

FIGURE 3.—CORRELATION CHART, SOUTHEASTERN ALBERTA TO CENTRAL ALBERTA PLAINS
Column on left shows approximate equivalents in northern U. S. Plains region; not to scale.

Southeastern Alberta Section to Central Alberta Plains Section

The recognition of the equivalents of the Whitemud and Battle formations of southeastern Alberta within the Edmonton formation of the central Alberta plains was first made by Sanderson (1931) and later supported by the writer (1932a, p. 127, 134). This correlation indicates that at least an upper portion of the Edmonton formation is equivalent in age to the Fox Hills stage, because the Whitemud and Battle formations are almost certainly the northern extensions of the upper Fox Hills. So much of the Edmonton formation lies below the Whitemud equivalent that it seems likely that the lower Fox Hills and a portion of the upper Pierre are also represented. To this correlation of the Edmonton formation as equivalent to

uppermost Pierre, and Fox Hills, must now be added the equation of the uppermost Edmonton beds with the Lance, as revealed by Sternberg's discovery (1947) of the *Triceratops* fauna in this portion of the formation. In many places these Lance equivalents of the Edmonton, and even older beds, have been removed by pre-Paskapoo erosion, but enough remnants exist to show that the *Triceratops* fauna was widely distributed and very characteristic. This Lance correlation of the uppermost Edmonton raises some curious problems on the relationships of the Battle-Frenchman unconformity of southwestern Saskatchewan to the Edmonton-Paskapoo unconformity of central Alberta. Obviously the two breaks are not identical, because in the first the *Triceratops* fauna occurs above the unconformity, and in the second the same fauna is present below. It is possible that an unconformity occurs within the Edmonton formation, between the Lance equivalent and the underlying portion, the Battle and Whitemud equivalents or lower, but such a break has not been detected and must in any case represent a relatively short time interval. The important interval in the Central Alberta Plains Section is that occurring between the uppermost Edmonton, equivalent to the Lance stage in places, and the Paskapoo formation, which seems to be exclusively Late Paleocene. That this interval represents Early and Middle Paleocene time seems reasonably well established, but its relationship to the Lance stage is still not entirely clear. Two possibilities exist. In one interpretation the central Alberta unconformity is an entirely different time break from that occurring in southwestern Saskatchewan and southeastern Alberta, and starting later in time, also terminated later. Another possible view would regard the unconformity as continuous spatially, but with different time limits in different places. According to the latter hypothesis deposition in the southeast ceased and erosion commenced at about the close of the Fox Hills stage. Meanwhile deposition in central Alberta continued on into the Lance stage. Then Lance sediments spread eastward across the irregular erosion surface of Battle or older strata, and about the same time or very shortly thereafter deposition ceased in central Alberta, to be followed in very late Cretaceous or Paleocene time by an interval of rather pronounced erosion. This is the view followed in preparing the diagram (Fig. 3) which shows the transitional relationships of the unconformity across the area of east-central Alberta now occupied by older formations. Correlation of the Paskapoo formation with the Ravenscrag of the southeast is still not definite, as mammalian faunas are lacking from the Ravenscrag. If, however, the writer's views as to the correlation of the Alberta Ravenscrag with that of Saskatchewan are correct, the Alberta Ravenscrag is probably approximately equal in age to the Paskapoo, that is, exclusively Late Paleocene.

Southwestern Alberta Section to Central Alberta Plains Section

The general correlation of the St. Mary River formation of southwestern Alberta with the Edmonton formation of the central Alberta plains is now usually recognized. The interfingering of the two lithological types, described by the writer along Little Bow river, has been confirmed by later workers. Uncertainty still exists, however, as to the upper boundary, that is, the horizon in the southwest which is equivalent to the top of the Edmonton in central Alberta. Outcrops are scarce along Little Bow river and adjacent streams in that portion of the section that is critical to the problem. As pointed out previously, the Paleocene molluscan fauna and the more recently studied Fort Union plants of the Willow Creek formation are confined to the upper part. On Little Bow river the writer (1932a, p. 128) found dinosaur bones in beds of essentially Edmonton lithological characteristics, and placed the Cretaceous-Tertiary boundary above this occurrence. He was unable to establish, however the relationships of this portion of the section to the St. Mary River-Willow Creek contact farther to the southwest. As a working arrangement, the writer drew the Cretaceous-Tertiary boundary at the St. Mary River-Willow Creek contact. W. A. Bell (personal communication) favors the location of this boundary within the Willow Creek formation itself, on the grounds that the Paleocene flora and fauna have been found only in the upper portion of the formation, and on the presence of conglomeratic

sandstone near the middle. Similar conglomeratic sandstone apparently also occurs on Little Bow river, above the highest dinosaur-bearing beds discovered by the writer. It is further assumed that these conglomeratic sandstones are the equivalents of the basal Paskapoo sandstone of the Central Alberta Plains Section. To this correlation, which would place a large lower portion of the Willow Creek formation within the Cretaceous, making it equivalent to uppermost Edmonton beds farther north, the writer has no objection, but such an interpretation is not as yet completely proved. The presence of Lance equivalents in the St. Mary River formation leaves little time available for deposition of several hundred feet of Late Cretaceous Willow Creek beds. At least the upper portion of the Willow Creek formation is still equated with the lower part of the Paskapoo formation. The attempt by some workers to align the Willow Creek-Porcupine Hills contact with the Cretaceous-Tertiary boundary and the Edmonton-Paskapoo contact is obviously opposed by both paleontological and stratigraphical evidence. Because of the scarcity of outcrops in the critical area, the main hope for future solution lies in the discovery of diagnostic fossils in the lower portion of the Willow Creek formation.

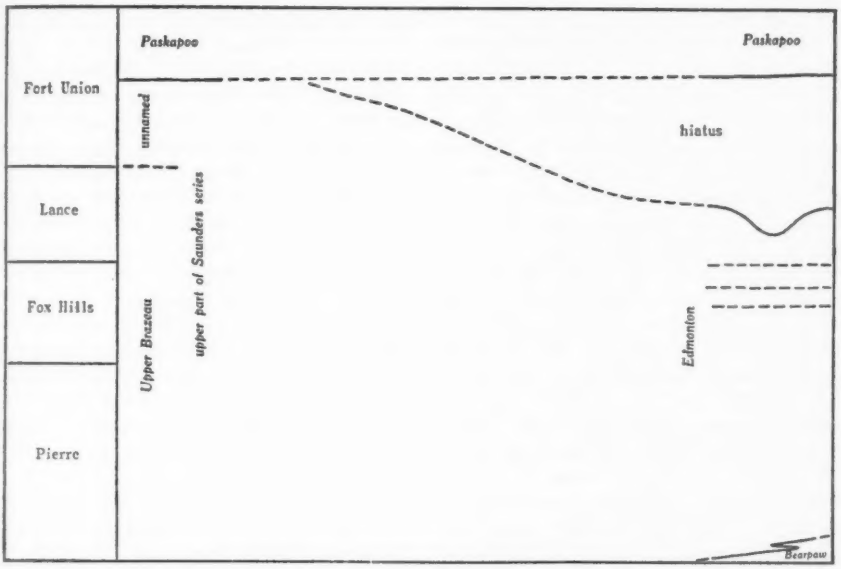

FIGURE 4.—CORRELATION CHART, CENTRAL ALBERTA PLAINS TO CENTRAL ALBERTA FOOTHILLS
Column on left shows approximate equivalents in northern U. S. Plains region; not to scale.

Central Alberta Plains Section to Central Alberta Foothills Section

Westward from the central Alberta plains (Fig. 4), the Edmonton-Paskapoo contact disappears beneath a thick series of Paskapoo strata occupying the axial portion of the Alberta syncline. On the west limb of this broad trough, where equivalent strata emerge again, the sandstone that is taken to represent the base of the Paskapoo overlies strata that do not resemble the Edmonton beds lithologically, but display the appearance of the Paskapoo strata, especially those well above the basal sandstone. These sub-Paskapoo strata consist of alternating sandstones and shales, a variegated sequence of fresh-water deposition. They pass downward with no important change or break into the great non-marine sequence of the foothills, to which the names Saunders series and Foothills series have been applied. If the massive,

conglomeratic sandstone of the west side of the Alberta syncline is truly the equivalent of the basal Paskapoo sandstone on the east, and this seems a reasonable interpretation, then on purely stratigraphical grounds the beds below should be equated with the upper Edmonton, in spite of the lithological differences (Rutherford 1928, p. 24; 1947). According to Rutherford the difference between the sub-Paskapoo strata of the west and the upper part of the Edmonton formation of the east is merely the result of different conditions of sedimentation; deltaic and partly brackish water on the east, purely fresh-water lakes and streams on the west. The findings of the paleontologists have long been opposed to this view; these upper strata of the Saunders series contain Paleocene plants, Fort Union-type fresh-water mollusks, and occasionally mammal teeth of definitely Paleocene age. To reconcile the apparent contradiction of stratigraphical and paleontological evidence, Russell (1932a, p. 147) suggested that the pre-Paskapoo Tertiary of the Central Alberta Foothills Section represented a portion of the time of the Edmonton-Paskapoo unconformity of the east, in other words Early and Middle Paleocene time, and probably also a portion of the Lance stage in the Late Cretaceous. No serious objection has been raised to this explanation, and the extension of the Paleocene lower limit to a much lower position in the series, as established by Bell on paleobotanical grounds (Lang, 1947, p. 32) urgently requires some such reconciliation of the two lines of evidence. The discovery of an incomplete tooth of the mammalian genus *Pantolambda* far down in the Saunders series on North Saskatchewan river (Russell, 1948) almost proves this interpretation; it shows that Middle Paleocene strata are present here, whereas, as we have seen, these are apparently missing from the Central Alberta Plains Section.

Summary

Conclusions established

The gradual nature of the Bearpaw Sea withdrawal is now generally recognized, and deltaic sandstones on the west have been correlated with marine shales on the east. The magnitude of this transition, however, is probably not fully appreciated by all workers. It is now agreed that the St. Mary River formation of the Southwestern Alberta Section and the Edmonton formation of the Central Alberta Plains Section are essentially equivalent, and that the lithological transition occurs between Oldman and Little Bow rivers. That the Willow Creek formation, which overlies the St. Mary River, is in part at least of Paleocene age has been confirmed by the paleobotanical findings of Bell. The Late Paleocene dating of the Paskapoo formation, with consequent assignment of the Edmonton-Paskapoo unconformity to Early and Middle Paleocene time, seems to be rather generally accepted. Paleobotany has supported the writer's conclusions in the central Alberta foothills, where the equivalence of a great thickness of pre-Paskapoo Saunders strata with the Paleocene has been confirmed. Further support is given by the discovery of a typical Middle Paleocene mammal fossil within this portion of the series. A Paleocene dating for this great thickness of Saunders strata within the disturbed belt further supports the writer's repeatedly expressed view that the major Rocky Mountain orogeny in the Canadian Cordillera was of post-Paleocene age.

Conclusions requiring further confirmation

The exclusively Paleocene age of the Ravenscrag formation in southeastern Alberta is denied by Furnival; who regards the lower portion of the post-Battle strata there as equivalent to his Frenchman formation, i.e., Lance, of southwestern Saskatchewan. Should Furnival's view prove correct, it would also invalidate the writer's correlation of the Alberta Ravenscrag with only the upper part of the type Ravenscrag of Saskatchewan and with the Paskapoo formation of the Central Alberta Plains Section. The Tertiary age of the Willow Creek formation, never positively maintained for the lower part by the writer, has been seriously questioned, because of the absence of diagnostic fossils within this portion of the formation and the apparent presence of an important stratigraphical break within the Willow Creek formation. The relationships of the Battle-Frenchman unconformity in Saskatchewan to the Edmonton-Paskapoo unconformity in central Alberta are not satisfactorily established as yet. The writer's

view of physical continuity of the break, which nevertheless varied in its time limits from east to west, is made somewhat less probable by the narrowing of the time gap on the west, owing to the discovery of Lance equivalents in the uppermost Edmonton. This makes it necessary to postulate a more pronounced shift in time from east to west. Within the central Alberta foothills, the presence of Early Paleocene equivalents remains to be established, but the extension by paleobotany of the Paleocene dating to a portion of the Saunders series well below that regarded previously as Tertiary allows adequate thickness for a complete Paleocene section.

Conclusions disproved

The only major and positive change required in the writer's previous conclusions is the extension of the Edmonton formation, which was definitely correlated with upper Pierre and Fox Hills, to include a portion equivalent to part of the Lance stage. In the original discussion, the presence of Lance equivalents in the uppermost Edmonton was considered but thought doubtful. C. M. Sternberg has established the presence of Lance equivalents in the uppermost portion of the Edmonton formation, where this has survived pre-Paskapoo erosion.

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TORONTO, ONT., CANADA

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NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION AND HIGHWAY DEPARTMENT

NEW HAMPSHIRE
LOVEWELL MOUNTAIN QUADRANGLE

(Claremont) (Sunapee) (Mount Kearsarge)

72°15' 72°00' 43°15'

350 000 FEET

270 000 FEET

A A'

B B'

10' 5'

(Bellows Falls) (Hillsboro)

SULLIVAN CO. MERRIMACK CO. HILLSBOROUGH CO.

Cheshire Co.

Lempster Mtn. East Lempster East Acworth Bridle Hill Cross Mtn. Baker Cor. Meetinghouse Hill Hilliard Pd. Kidder Pd. Hubbard Hill Wildcat Hill Wilkins Hill Wright Hill Beckman Hill Grassy Pond Hill Lovewell Mtn. Farnsworth Hill Miller Lake Ashuelot Pond Gregg Hill Hubbard Hill Highland Lake Goshen Hill Kingsbury Hill Bald Hill Black Hill Mt. Nancy Mtn. Ragged Mtn. Saddle Hill Stoddard Hill Watatic Mtn. East Washington Pd. Washington Pd. Hunt Hill Lovewell Pd.

LATE DEVONIAN?

New Hampshire Magma Series

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LOWER DEVONIAN

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NG
AY DEPARTMENT

NEW HAMPSHIRE
LOVEWELL MOUNTAIN QUADRANGLE

(Sunapee)
(Mount Kearsarge)
72°00'
43°15'
A
A'
10'
5'
WASHINGTON
STODDARD
Dodge Hill
Hubbard Hill
Sawed Hill
Robinson Hill
Parker Hill
Sharp Hill
Lovell Hill
Woodbury Hill
Brooks Hill
Farnsworth Hill
Hunkler Mtn
Silver Mtn
Beech Hill
Millen Lake
Ashuelot Pond
Highland Lake
Island
Kidder Pond
Long Pond
Center Pond
Black Pond
White Pond
Island Pond
Sugar River
Ashuelot River
Warren Brook
Mill Brook
Mill Pond
Meeting House Brook
Mountain Brook
Mine Brook
Mine Pond
Old Mine
Old Mine Pit
Millbrook Pit
Kingsbury Pit
Woodward Pit
Hubbard Pit
Sullivan Co.
Cheshire Co.
Hillsboro Co.
(Hillsboro)
Adams Cliff
Oakhill
Dale Hill
Mt. Brooks
Mountain View Camp
Camp
Cabin
Old Cellar Hole
Old Cellar
Old Mill
Old Dam
Old Pit
Old Quarry
Old Road
Old Trail
Old Tower
Old Well
Old Woods
Old Woods Road
Old Woods Trail
Old Woods Trail
Old Woods Trail

SECTION A-A′

SECTION B-B′

SECTION C-C′

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE LOVEWELL MOUNTAIN QUADRANGLE, NEW HAMPSHIRE

Topographic base by U. S. Geological Survey. Surveyed in Cooperation with the State of New Hampshire.

Scale 1/62500

1 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet

Datum is mean sea level

APPROXIMATE MEAN DECLINATION, 1950

Geology by M. T. Heald assisted by J. B. Carney and J. R. Williams. Directed by Marland P. Billings. Geology survey 1946 and 1947, with the aid of grants from the R. W. Sayles Fund of Harvard University. Published in 1950.

SECTION A-A'

SECTION B-B'

SECTION C-C'

LOWER

Littleton formation
Hubbard Hill member

(Chiefly mica-quartz schist, sillimanite schist, pseudo-sillimanite schist (Dlh)*; quartz conglomerate and quartzite* (Dlq)*; interbedded lime-silicate granulite and schist* (Dlf)*; amphibolite* (Dlv)*.)*

CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

Indefinite as sharp contact is lacking

Strike and dip of bedding

Strike of vertical beds

Strike and dip of overturned beds

Strike and dip of foliation

Strike of vertical foliation

Mines, prospects, and quarries, mostly abandoned

E SECTIONS OF THE LOVEWELL MOUNTAIN QUADRANGLE, NEW HAMPSHIRE

Scale 1/62500

1 0 1 2 3 4 Miles
5000 0 5000 10000 15000 20000 Feet
1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet
Datum is mean sea level

Geology by M. T. Heald assisted by J. B. Carney and J. R. Williams. Directed by Marland P. Billings. Geology survey 1946 and 1947, with the aid of grants from the R. W. Sayles Fund of Harvard University. Published in 1950.

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Fig
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2.

3.

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STRUCTURE AND PETROLOGY OF THE LOVEWELL MOUNTAIN QUADRANGLE, NEW HAMPSHIRE

By Milton T. Heald

CONTENTS

	Page
Abstract	44
Introduction	45
Acknowledgments	45
General lithologic features	45
Littleton formation	45
General statement	45
Hubbard Hill member	46
General statement	46
Schist	47
Lime-silicate granulite and associated schist	48
Quartz conglomerate and quartzite	48
Gneisses	50
Amphibolite	50
May Pond member	50
General statement	50
Lithologic characteristics	50
Dakin Hill member	54
General statement	54
Porphyroblastic orthoclase gneiss	54
Biotite gneiss	54
Pyritiferous gneiss	55
Thickness	55
Age	55
Plutonic rocks	55
General statement	55
Bethlehem gneiss	55
Kinsman quartz monzonite	58
General statement	58
Bacon Ledge pluton	58
Huntley Mountain pluton	61
Minor bodies	61
Concord granite	61
Miscellaneous dike rocks	61
Age of the New Hampshire magma series	63
Mineralogy	63
General statement	63
Potash feldspar	63
Plagioclase	63
Amphibole	64
Pyroxene	64
Muscovite	64
Biotite	65
Garnet	65
Cordierite	65
Distribution of constituents in garnet, biotite, and cordierite	67
Heavy minerals	68
Structural Features	69
General statement	69
Metasedimentary rocks	69
General statement	69
Hubbard Hill member	69
May Pond member	70
Dakin Hill member	70
Bethlehem gneiss	70
Mt. Clough pluton	70
Minor bodies of Bethlehem gneiss	72
Kinsman quartz monzonite	72
Bacon Ledge pluton	72
Huntley Mountain pluton	73
Metamorphism of the metasedimentary rocks	74
General statement	74
Progressive metamorphism	74
Retrograde metamorphism	75
Origin of paragneisses	75
Changes in chemical composition	78
General statement	78
Hubbard Hill member	78
May Pond and Dakin Hill members	79
Summary of metamorphism	79
Origin of the plutonic rocks	79
General theory	79
Evidence of intrusion	79
General statement	79
Kinsman quartz monzonite	79
Bethlehem gneiss	81
Evidence of replacement	82
Kinsman quartz monzonite	82
Bethlehem gneiss	82
Conclusions	82
Syntectonic nature of intrusion	82
Metamorphism of the plutonic rocks	82
Origin of the large feldspar crystals in the Kinsman quartz monzonite	84
Origin of the magma	86
General statement	86
Differentiation of basaltic magma	86
Assimilation of sediments by basaltic magma	86
"Pure melting" of older granitic rocks	86
Solution of beds at a comparatively shallow depth	86
Conclusions	87
Summary of geologic history	87
References cited	88

ILLUSTRATIONS

Figure	Page
1. Index map	46
2. Columnar section of the metasedimentary rocks	47
3. Plutonic rocks in the eastern part of the Bellows Falls quadrangle	56
4. Relation between axial angle and $K_2O + Na_2O - CaO$ for cordierite	67
5. Distribution of FeO, MgO, and Al_2O_3 in garnet, biotite, and cordierite	67

Figure Page
6. Contact between inclusion of schist and Kinsman quartz monzonite..... 72
7. Cross-cutting stringer of Kinsman quartz monzonite..... 73
8. Metamorphic subzones in the Lovewell Mountain quadrangle..... 74
9. Projection of the analyzed paragneisses.... 78
10. Contact between schist and Kinsman quartz monzonite..... 80
11. Pods of Kinsman quartz monzonite in schist 80
12. Inclusion of schist in the Kinzman quartz monzonite..... 81
13. Contact between schist and Kinsman quartz monzonite..... 81
14. Composition of garnets from granites..... 83
15. Orthoclase-plagioclase equilibrium diagram 85

Plate Facing page
1. Geologic map and structure sections of the Lovewell Mountain quadrangle, New Hampshire..... 43
2. Metasedimentary rocks..... 48
3. Folds and stretched pebbles in the Hubbard Hill member..... 49
4. Tectonic map of the Lovewell Mountain quadrangle, New Hampshire..... 88

Table Page
1. Section of interbedded sillimanite schists and mica-quartz schists of the Hubbard Hill member..... 48
2. Estimated modes of the Hubbard Hill member..... 49
3. Measured section of lime-silicate beds and associated rocks..... 50

Table Page
4. Estimated modes of the May Pond Member. 51
5. Estimated modes of the Dakin Hill member. 52
6. Estimated modes of the Bethlehem gneiss and associated schists..... 57
7. Estimated modes of the Kinsman quartz monzonite and schist inclusions, Bacon Ledge pluton..... 59
8. Estimated modes of the Kinsman quartz monzonite and miscellaneous dike rocks.. 62
9. Chemical analyses of minerals from the Lovewell Mountain quadrangle..... 64
10. Optically positive cordierites..... 66
11. Distribution of MnO and TiO_2 in garnet, biotite, and cordierite..... 67
12. Heavy minerals from metasedimentary and plutonic rocks in weight per cent of bulk sample..... 68
13. Attitude of observed contacts between the Bethlehem gneiss and schist..... 71
14. Analyses of rocks containing carbon..... 76
15. Chemical analyses of metamorphosed argillaceous rocks from the Littleton formation..... 77
16. Characteristics of the Kinsman quartz monzonite and the Bethlehem gneiss..... 80
17. Chemical analyses of Kinsman quartz monzonite..... 84
18. Composition of the least amount of material which can be added to average plateau basalt to produce the Kinsman quartz monzonite..... 86
19. Average modes of banded gneiss and Bethlehem gneiss..... 87

Abstract

The Lovewell Mountain quadrangle in southwestern New Hampshire is underlain by metasedimentary rocks of the lower Devonian Littleton formation and plutonic rocks of the New Hampshire magma series, which is probably of late Devonian age,

The lower portion of the Littleton formation in the area consists mainly of well-bedded sillimanite schist, pseudo-sillimanite schist, and mica-quartz schist, whereas the upper portion consists of poorly banded gneisses. The thickness of the Littleton formation exposed in the area is probably of the order of 20,000 feet. The plutonic rocks range from quartz monzonite to quartz diorite but granodiorite predominates.

The area is on the eastern flank of a large anticline. The foliation in the metasedimentary rocks is mainly of the axial plane type and is parallel to bedding except at the noses of folds. The plutonic bodies are generally concordant although some parts of the bodies cross-cut several thousand feet of the adjacent metasedimentary rocks. Small scale discordant relationships are common along the contacts of the Kinsman quartz monzonite.

The metasedimentary rocks have been subjected to high-grade metamorphism. Orthoclase formed at the expense of muscovite in a large part of the area in response to the high temperatures which prevailed during the peak of metamorphism, but the orthoclase altered to muscovite and quartz under declining intensity conditions during the final stages of metamorphism in most of the eastern two-thirds of the quadrangle. Changes in chemical composition during metamorphism were apparently not important, although potash was probably added to the pseudo-sillimanite schists and soda and lime were added to a few of the paragneisses.

Most of the bodies of plutonic rocks are intrusive, but the gradational nature of the contacts in a few places suggests that some of the plutonic rocks may have undergone little movement. Definite conclusions regarding the source of the granitic material could not be drawn. Magma rising from great depth may have entered certain favorable horizons to form large sill-like masses. The mineralogical and textural similarity between the plutonic rocks and some of the metamorphic rocks suggests that the granitic material may have originated at a comparatively shallow depth by the solution or partial solution of favorable beds. The mobilized material may have moved along approximately the

same stratigraphic horizon at which it originated to form the intrusive sheets.

Introduction

The Lovewell Mountain quadrangle is located in the southwestern part of New Hampshire (Fig. 1). Early investigations of the bedrock geology in the area were conducted by C. T. Jackson (1844) and C. H. Hitchcock (1875; 1877; 1878). Daly (1897) made brief study of the Kinsman quartz monzonite in the eastern part of the quadrangle. Conant (1935) reported optically positive cordierite and Meyers (1948) described green lazulite from the area. Modern mapping in western New Hampshire has been completed north of the area except in the Sunapee quadrangle where C. A. Chapman is now conducting field work. South of the Lovewell Mountain quadrangle Katharine Fowler-Billings (1949) has completed mapping through the Monadnock quadrangle to the Massachusetts line.

The field work occupied 31 weeks during the summers of 1946 and 1947. The geology was plotted on the 1942 edition of the United States Geological Survey topographic sheet enlarged to a scale of 3 inches to a mile. Most locations were made by inspection or by pace and compass traverses but an altimeter was used on the steeper slopes. Some points were located by means of aerial photographs taken in 1939 by the United States Department of Agriculture. About 12 days were spent making pace and compass maps (scale of 750 feet to the inch) along critical contacts.

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General Lithologic Features

The metasedimentary rocks (Fig. 2) in the Lovewell Mountain quadrangle belong to the Devonian Littleton formation and comprise about half of the area. Largely schists and gneisses, they are in the high-grade metamorphic zone.

The plutonic rocks belong to the late Devonian (?) New Hampshire magma series, which is represented in this area by the Bethlehem gneiss, the Kinsman quartz monzonite, and the Concord granite. Post-metamorphism quartz veins and light-colored granitic dikes are common, especially in the bodies of Kinsman quartz monzonite. Large bodies of pegmatite are numerous in the western part of the quadrangle and small veins of pegmatite are abundant throughout the plutonic bodies.

Littleton Formation

General Statement

In the Lovewell Mountain quadrangle the Littleton formation has been subdivided into three members: the Hubbard Hill member, the May Pond member, and the Dakin Hill member. The Hubbard Hill member is the lowermost portion of the Littleton formation in the area. The base of this member lies about 5000 feet above the base of that portion of the Littleton formation which is exposed to the west in the Bellows Falls quadrangle (Kruger, 1946, p. 186). The Hubbard Hill member is exposed in the southwestern and northwestern portions of the quadrangle. The May Pond member lies stratigraphically above and to the east of the Hubbard Hill member. The uppermost portion of the Littleton formation, the Dakin Hill member, occupies a large area extending from the southern boundary to the northern boundary of the quadrangle in the central portion of the area.

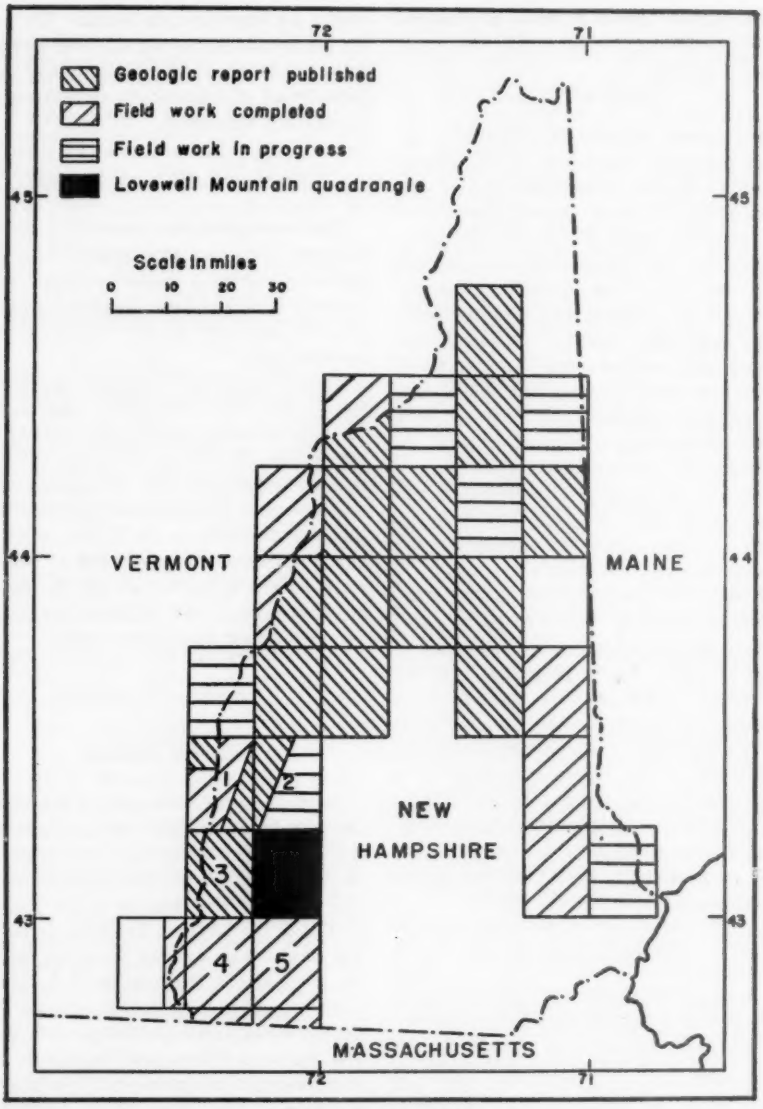

FIGURE 1.—INDEX MAP

Areas adjacent to the Lovewell Mountain quadrangle in which mapping has been recently completed or field work is in progress are: 1 = Claremont quadrangle (R. W. Chapman and C. A. Chapman); 2 = Sunapee quadrangle (C. A. Chapman); 3 = Bellows Falls quadrangle (F. C. Kruger, 1946); 4 = Keene area (G. E. Moore, 1949); 5 = Monadnock region (K. Fowler—Billings, 1949).

Hubbard Hill Member

General Statement.—The Hubbard Hill member in the northwestern part of the area is composed largely of interbedded sillimanite schists and mica-quartz schists.[1] Discontinuous beds of lime-silicate granulite and quartz conglomer-

[1] In accordance with the classification used by Billings (1937, p. 491), rocks composed mainly of mica and quartz with a little feldspar are desig-

ate are exposed in the west-central part of the area, and small masses of amphibolite are found in a few places. In addition to sillimanite schist and mica-quartz schist, banded gneiss occurs locally in the Hubbard Hill member in the southwestern part of the area.

The chemical composition of the minerals of these rocks is considered in a later section of the paper.

Schist.—Interbedded sillimanite schist and mica-quartz schist comprise most of the Hubbard Hill member in the northwestern part of the area. Quartz-mica schist and quartz-feldspar-mica schist occur locally. Individual beds are only a few inches thick, but are continuous over considerable distances (Pl. 2, fig. 1). A representative measured section of the schists is given in Table 1. In the southwestern part of the area bedding is not as conspicuous as in the northwestern portion.

The schists are gray and medium-grained to coarse-grained. Modes are given in Table 2, schistosity and in a few places show a slight linear parallelism. Most of the porphyroblasts have been drawn out into augen and are partially altered to muscovite. Pseudo-sillimanite schist occurs locally where the sillimanite has largely altered to muscovite. Although the sillimanite has been completely replaced in some of the schists, small remnants of sillimanite are generally found in the core of the muscovite pseudomorphs. Accessory tourmaline is slightly more abundant in the pseudo-sillimanite schists than in the other types. The pseudo-sillimanite schists near the northern boundary of the quadrangle are moderately contorted and contain small lenses of quartz and feldspar. In this area, large cross-muscovite flakes are conspicuous in the matrix as well as in the pseudomorphs. The fact that the muscovite is only slightly deformed indicates that it formed after most of the deformation.

columns 1 to 5. The sillimanite in the sillimanite schists generally occurs in porphyroblasts which are 2–5 cm. long and 5 mm. thick. The porphyroblasts generally lie in the plane of

AGE	FORMATION	MEMBER	SYMBOL	COLUMNAR SECTION	LITHOLOGY	THICKNESS IN FEET
LOWER DEVONIAN	LITTLETON FORMATION	OAKIN HILL	Dlo		Porphyroblastic orthoclase gneiss, biotite-quartz gneiss, biotite-sillimanite gneiss, and pyritiferous gneiss.	10,000(?)
		MAY POND	Dlm		Orthoclase gneiss and biotite gneiss.	0-3,000(?)
		HUBBARD HILL	Dlh Dlq Dll Dlh Dlv		Mica-quartz schist, quartz-mica schist, quartz-feldspar-mica schist, sillimanite schist, pseudo-sillimanite schist, and gneiss (Dlh); quartz conglomerate and quartzite (Dlq); diopside granulite, actinolite-diopside granulite, and purplish biotite schist (Dll); amphibolite (Dlv).	6,000

FIGURE 2.—COLUMNAR SECTION OF THE METASEDIMENTARY ROCKS

nated as follows: mica schist, less than 60 per cent quartz and feldspar; mica-quartz schist, 60 to 80 per cent quartz and feldspar; quartz-mica schist and quartzite, over 80 per cent quartz and feldspar.

TABLE 1.—SECTION OF INTERBEDDED SILLIMANITE SCHISTS AND MICA-QUARTZ SCHISTS OF THE HUBBARD HILL MEMBER

Summit of Silver Mountain, Lempster

	Thickness (inches)
Mica-quartz schist	1.0
Sillimanite schist	0.5
Mica-quartz schist	1.1
Sillimanite schist	0.8
Mica-quartz schist	1.0
Sillimanite schist	0.1
Mica-quartz schist	1.0
Sillimanite schist	7.0
Mica-quartz schist	0.1
Sillimanite schist	5.5
Mica-quartz schist	0.5
Sillimanite schist	0.5
Mica-quartz schist	0.3
Sillimanite schist	2.0
Mica-quartz schist	0.3
Sillimanite schist	3.0
Mica-quartz schist	1.5
Sillimanite schist	1.5
Mica-quartz schist	1.0
Sillimanite schist	0.3
Mica-quartz schist	0.8
Sillimanite schist	11.0
Mica-quartz schist	0.5
Sillimanite schist	0.3
Mica-quartz schist	0.3
Sillimanite schist	2.3
Mica-quartz schist	2.0
Sillimanite schist	6.0
Mica-quartz schist	0.3
Sillimanite schist with a few 0.3 inch mica-quartz schist beds	17.0
Mica-quartz schist	1.0
Sillimanite schist	1.3
Mica-quartz schist	0.3
Sillimanite schist	2.0
Mica-quartz schist	0.3
Sillimanite schist	2.0
Mica-quartz schist	0.3
Sillimanite schist with a few 0.3 inch mica-quartz schist beds	12.0
Sillimanite schist	0.5
Mica-quartz schist	0.5
Sillimanite schist	1.5
Mica-quartz schist	1.0
Sillimanite schist	0.5
Mica-quartz schist	0.8
Total	93.5

Sillimanite schist.....83 per cent
Mica-quartz schists.....17 per cent

Lime-silicate granulite[2] *and associated schist.*—Beds of lime-silicate granulite occur discontinuously at one horizon in the Hubbard Hill member between Gee Mill and the western knoll of Marlow Hill in Marlow (Pl. 1). A measured section of the lime-silicate beds and associated schists is given in Table 3. The granulites are medium-grained and light gray or pale green. A few of the beds are pyritiferous and have a rusty-brown weathered surface. The lime-silicate rocks were probably derived from impure dolomitic limestones.

Microscopic study shows that the texture is essentially granoblastic and that the grain size ranges from 0.1 to 1.0 mm. Essential minerals are actinolite, diopside, plagioclase, microcline, and quartz. The plagioclase ranges from bytownite (An_{67}) to anorthite (An_{93}) in different specimens. Common accessories are garnet, biotite, pyrite, and sphene. Modes are given in Table 2, columns 8, 9, and 10.

Purplish biotite schists and pyritiferous mica schists are interbedded with the lime-silicate granulites. The schists are fine-grained to medium-grained and are composed mainly of quartz, anorthite, biotite, and garnet. A mode of the purplish biotite schist is given in Table 2, column 11. Originally these schists were probably calcareous shales.

Quartz conglomerate and quartzite.—Quartz conglomerates and quartzites overlie the lime-silicate beds. The quartz conglomerates crop out almost continuously from Gee Mill to the southwestern slope of Marlow Hill (Pl. 1). Discontinuous lenses of quartz conglomerate and quartzite, which are too small to be shown on the geologic map, are common between Marlow Hill and the western boundary of the quadrangle. Most of the pebbles in the conglomerate are composed of quartz and are 1 inch to 2 feet long. They are highly drawn out and have axial lengths ranging from 1:2:4 to 1:3:16. The matrix generally contains more than 80 per cent quartz, but locally it is rich in muscovite. Biotite, oligoclase, and microcline occur in minor amounts with accessory pyrite, sphene, zircon, and garnet. A mode is given in Table 2, column 6. Only one outcrop of relatively pure quartzite has been observed in

[2] Granulite, as used in this paper, refers to quartzo-feldspathic metamorphic rock without conspicuous schistosity.

FIGURE 1. HUBBARD HILL MEMBER
Interbedded mica-quartz schist and sillimanite schist.

FIGURE 2. MAY POND MEMBER
Crudely banded biotite gneiss. Light-colored lenses are composed of quartz, muscovite, and plagioclase.

FIGURE 3. DAKIN HILL MEMBER
Orthoclase gneiss. The light-colored clots are largely orthoclase with a few per cent of quartz.

METASEDIMENTARY ROCKS

FIGURE 1. FOLDS IN SCHIST

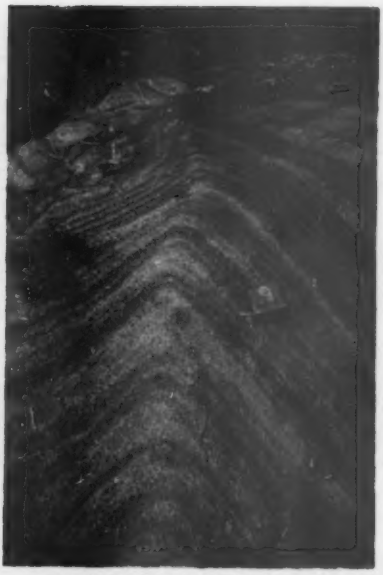

FIGURE 2. FOLDS IN SCHIST

FIGURE 3. FOLDS IN QUARTZ VEINS AND QUARTZITE

The bed of quartzite is shown to the left of the hammer. The quartz veins, which occur in micaceous beds, are more highly contorted than the quartzite.

FIGURE 4. STRETCHED PEBBLES IN QUARTZ CONGLOMERATE

FOLDS AND STRETCHED PEBBLES IN THE HUBBARD HILL MEMBER

Photographs by J. R. Williams

TABLE 2.—ESTIMATED MODES OF THE HUBBARD HILL MEMBER

	Schists					Quartzites		Lime-silicate granulites and associated schists				Amphibolites
	1	2	3	4	5	6	7	8	9	10	11	12
Number of thin sections	1	3	1	5	2	1	1	1	2	1	2	1
Quartz	79	53	55	51	59	88	94		3	10	40	5
Microcline							3	38	32			
Oligoclase	5	13	28	14	4	6						
Andesine												51
Bytownite									5			
Anorthite								20		48	33	
Biotite	12	22	12	21	18	6	tr	2		tr	22	3
Muscovite		9		5	13	tr	2				tr	
Chlorite		tr					tr					
Hornblende												40
Actinolite								tr	37	39		
Diopside								40	23	tr		
Sillimanite	tr	tr		8	1							tr
Garnet	4	2	5	1	4	tr	tr			1	4	
Magnetite	tr	1	tr	tr	tr	tr	tr	tr			tr	tr
Tourmaline	tr	tr		tr	1				tr	tr		
Pyrite								tr	tr	2	1	
Apatite	tr	tr	tr	tr	tr	tr					tr	tr
Zircon	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	tr	
Sphene							tr	tr	tr	tr		
Graphite				tr					tr	tr		
Zoisite							1					
Rutile							tr					
Epidote												1
Per cent of anorthite in plagioclase	25	27	28	27	28	28		93	87	90	92	35
Grain size in mm.	0.1–1.0	0.1–3.0	0.2–1.0	0.1–40	0.2–8	0.1–1.5	0.5–1.0	0.1–1.0	0.1–1.0	0.2–1.0	.05–1.0	0.2–1.0
Texture*	G	S	G	S	S	G	G	G	G	G	S	G

* G—Granulose; S—Schistose.

Schists

1. Quartz-mica schist.
2. Mica-quartz schist.
3. Quartz-feldspar-mica schist.
4. Sillimanite schist.
5. Pseudo-sillimanite schist.

Quartzites

6. Matrix of quartz conglomerate.
7. Quartzite.

Lime-silicate granulites and associated schists.

8. Diopside granulite.
9. Actinolite-diopside granulite.
10. Pyritiferous actinolite granulite.
11. Purplish biotite schist.

Amphibolites

12. Amphibolite.

TABLE 3.—MEASURED SECTION OF LIME-SILICATE BEDS AND ASSOCIATED ROCKS

Road cut at Gee Mill in Marlow

	Thickness (feet)
Top not exposed	
Actinolite-diopside granulite	0.4
Purplish mica-quartz schist	6.6
Actinolite-diopside granulite	1.0
Mica-quartz schist	1.8
Actinolite-diopside granulite	0.8
Interbedded pyritiferous mica-quartz schist and purplish biotite schist	6.8
Garnetiferous granulite	0.2
Interbedded pyritiferous mica-quartz schist and purplish biotite schist	4.2
Actinolite-diopside granulite	0.4
Concordant quartz vein	0.6
Actinolite-diopside granulite	2.5
Gap	1.7
Pyritiferous actinolite-diopside granulite	12.5
Interbedded pyritiferous actinolite-diopside granulite and mica schist	18.9
Pyritiferous mica-quartz schist	14.0
Pyritiferous actinolite-diopside granulite	12.3
Light-brown actinolite-diopside granulite	48.5
Pyritiferous actinolite-diopside granulite	2.8
Interbedded pyritiferous sillimanite schist and pyritiferous mica-quartz schist	12.8
Pegmatite	3.0
Pyritiferous sillimanite schist	1.2
Pegmatite	0.3
Sillimanite schist	1.8
Pegmatite	0.3
Sillimanite schist	0.8
Pegmatite	0.5
Mica-quartz schist and sillimanite schist	0.7
Pegmatite	2.4
Sillimanite schist with a few half-inch beds of mica-quartz schist	4.2
Interbedded pyritiferous mica-quartz schist and pyritiferous sillimanite schist	0.8
Sillimanite schist	1.2
Interbedded pyritiferous mica schist and sillimanite schist with a few beds of garnet granulite 3 mm. thick	3.0
Typical nonpyritiferous schists of the Hubbard Hill member	
Total (excluding pegmatites and quartz veins)	160.2

the area. This quartzite has a glassy appearance and contains 94 per cent quartz (Table 2, column 7).

Gneisses.—Gneisses are not common in the Hubbard Hill member, but they occur at a few localities in the southwestern part of the area. Well-formed lenses of light-colored minerals comprise 5–20 per cent of the rock. The lenses are several centimeters long and 5 to 10 mm. thick. They are composed of andesine and quartz with minor amounts of potash-feldspar, muscovite, and biotite. Most of the lenses are very coarse-grained, with andesine crystals as much as 2 cm. long. Because of the coarse-grained texture of the lenses, accurate modal determinations could not be made. It is estimated that the lenses contain 60–75 per cent andesine (An_{32}) and 25–40 per cent quartz. The dark layers in the gneiss are schistose and have the same mineralogical composition as ordinary mica schist. Some of the gneisses in the eastern portion of the Hubbard Hill member in Sullivan contain about 5 per cent potash-feldspar and are poorly foliated. These rocks are similar to the gneisses of the May Pond member which are described in a later section.

Amphibolite.—Small masses of amphibolite occur north of Gustin Pond in Marlow. The amphibolites are composed mainly of andesine and amphibole with minor amounts of biotite and quartz. A mode is given in Table 2, column 12. These rocks are similar to the amphibolites in the Littleton-Mooselauke area which are believed to be of igneous origin (Billings, 1937, p. 493).

May Pond Member

General statement.—The May Pond member lies stratigraphically above and to the east of the Hubbard Hill member. In the northern part of the quadrangle, the two members are separated by a sill of Kinsman quartz monzonite. The May Pond member is composed predominantly of banded gneiss although beds of well-stratified schist occur in a few places.

Lithologic characteristics.—The light-colored minerals in the gneiss are generally concentrated in short lenses rather than continuous layers (Pl. 2, fig. 2). In some of the gneiss, banding is vague and the light-colored constituents occur in irregular contorted stringers. In the well-foliated rocks, the lenses are thinner and more continuous than in the massive types.

The average lens is about 7 cm. long and 1.5 cm. thick, but some of the lenses reach a maximum of 20 cm. in length and 4 cm. in thickness. The degree of segregation of the light and dark minerals is highly variable. In some of the gneiss the light-colored aggregates are composed almost entirely of quartz, feldspar, and muscovite; the dark layers consist largely of biotite and garnet. More commonly the light-colored portions contain a moderate amount of biotite and the dark layers contain considerable quartz and feldspar. The ratio of light to dark layers ranges between wide limits. On the average, the light-colored bands constitute 10–20 per cent of the rock, but they may comprise as much as 30 per cent of the rock.

The dark layers are generally fairly schistose and are composed of biotite, quartz, oligoclase, garnet, and sillimanite. Muscovite does not occur in any of the unaltered rocks which contain orthoclase. Although the percentage of oligoclase in the dark layers is variable, it is unrelated to the amount of feldspar in the light-colored bands. The composition of the oligoclase is approximately the same in the light and dark layers. Sillimanite occurs in fibrous clots and as fine needles in biotite and quartz. Sillimanite porphyroblasts, 2–4 cm. long, are present in some of the gneisses near the contact with the Kinsman quartz monzonite.

In the west-central portion of the quadrangle, the light-colored bands in the gneiss are rich in orthoclase and poor in plagioclase. Some of the orthoclase forms large euhedral porphyroblasts 2–5 cm. long, but most of the orthoclase occurs in coarse-grained lenses with small amounts of quartz. Modes of the gneisses containing orthoclase are given in Table 4, columns 1 and 2. The mode of the orthoclase gneiss given in column 2 was determined from a crushed sample of a large block of gneiss.

In the northern part of the quadrangle, the light-colored portions of gneiss are composed mainly of quartz, muscovite, and oligoclase with only small amounts of orthoclase (Table 4, column 5). Some of the light-colored aggregates have the form of orthoclase crystals. It is believed that originally the light-colored portions of the gneiss consisted mostly of orthoclase which was replaced by muscovite, quartz, and oligoclase during a later period of retrograde metamorphism. The composition of the oligoclase in the pseudomorphs after orthoclase is approximately the same as that of the oligoclase in the enclosing rock. Modes of the gneiss occurring in the northern portion of the May Pond member are given in Table 4, columns 3 and 6.

TABLE 4.—ESTIMATED MODES OF THE MAY POND MEMBER

	1	2	3	4	5	6
Number of thin sections	2	*	3	3	*	2
Quartz	55	35	38	35	40	52
Potash feldspar	3	11				
Oligoclase	5	16	9	4	53	29
Biotite	24	27	35	40	2	12
Muscovite			4	3	5	1
Sillimanite	8	7	10	13		3
Garnet	5	4	4	5		3
Magnetite	tr	tr	tr	tr		tr
Pyrite			tr	tr		
Apatite	tr	tr	tr	tr		tr
Zircon	tr	tr	tr	tr		tr
Sphene			tr	tr		tr
Per cent of anorthite in oligoclase	27	28	30	30	30	28
Grain size in mm	0.2–30	0.2–10	0.05–4.0	0.05–2.0	0.05–4.0	0.03–2.0

* Mode based on thin sections and powdered samples.

1. Porphyroblastic orthoclase gneiss.
2. Banded orthoclase gneiss.
3. Biotite gneiss.
4. Dark portion of biotite gneiss.
5. Light-colored portion of biotite gneiss.
6. Quartz-oligoclase-biotite gneiss.

Metamorphosed concretions, similar to those described by Billings and others (1946, p. 264) in the Mt. Washington area are fairly common in the May Pond and Dakin Hill members. The concretions are ellipsoidal and range in size from a few inches to several feet. Megascopic study indicates that they are rich in quartz and garnet with green amphibole occurring in some specimens.

TABLE 5.—ESTIMATED MODES OF THE DAKIN HILL MEMBER

	Porphyroblastic orthoclase gneiss		Biotite gneiss							Pyritiferous gneiss				
	1	1a	2	3	4	4a	5	6	7	8	9	10	10a	11
Number of thin sections	2		5	3	1		1	1	1	3	2	2		1
Porphyroblasts														
Orthoclase	8	8.2												3
Garnet														2
Groundmass														
Quartz	36	40.5	58	46	37	35.2	39	44	40	58	48	40	39.0	46
Oligoclase	3	5.2	11		6	14.1	10	34	15	14	3	2	0.0	19
Oligoclase-andesine				4										
Biotite	28	25.1	22	35	37	29.3	24	17		15	32	6	6.1*	20
Muscovite	tr		3	2	3	3.0*	15	4	20	6	6	35	35.2	
Chlorite			tr	tr	1	1.0*			12	2		9	10.2	
Cordierite	2	2.0*									tr			
Sillimanite	15	11.5	3	7	7	7.3	tr		2	3	6	tr	1.2	7
Garnet	7	7.0*	3	6	7	9.3	11		10	tr	2			
Magnetite						0.1							0	
Ilmenite	1		tr	tr	2		1	1	1	tr	tr	1		tr
		0.5				0.5							0.7	
Pyrite		tr			tr	0.3				1	3	4	2.4	1
Tourmaline			tr											
Apatite	tr	0.3	tr	tr			tr	tr	tr	tr	tr	tr	0.3	tr
Zircon	tr		tr	tr	tr		tr	tr	tr	tr	tr	tr		tr
Sphene			tr	tr						tr		tr	0.2*	
Graphite			tr							1	tr	2	1.6	1
Zoisite	tr													
Myrmekite										tr				1
Melanterite (?)										tr		1	3.2	
Per cent of anorthite in plagioclase	28	28	26	29	30	30	28	29	30	28		28	16	27
Size of porphyroblasts in mm	20–50													10–30
Size of groundmass in mm	0.2–2.0		0 1–2.5	0.05–2.0	0.1–2.0		0.1–2.0	0.1–1.0	0.05–1.0	0.1–3.5	0.1–2.0	0.2–1.5		0.05–1.0

* Assumed to be equal to amount observed in thin section.

Porphyroblastic orthoclase gneiss.

* Assumed to be equal to amount observed in thin section.

Porphyroblastic orthoclase gneiss.

1. Porphyroblastic orthoclase gneiss.

1a. Porphyroblastic orthoclase gneiss, calculated from chemical analysis, specimen LM 6.

Biotite gneiss.

2. Biotite-quartz gneiss.
3. Biotite-sillimanite gneiss.
4. Biotite-sillimanite gneiss, specimen LM 125.

4a. Biotite-sillimanite gneiss, calculated from chemical analysis, specimen LM 125.

5. Biotite-muscovite gneiss.
6. Quartz-feldspar-biotite gneiss.
7. Chloritized biotite gneiss.

Pyritiferous gneiss.

8. Pyritiferous mica-quartz gneiss.
9. Pyritiferous biotite gneiss.
10. Pyritiferous muscovite gneiss.

10a. Pyritiferous muscovite gneiss, calculated from chemical analysis, specimen LM 358.

11. Pyritiferous orthoclase gneiss.

Dakin Hill Member

General Statement.—The Dakin Hill member is the uppermost portion of the Littleton formation exposed in the area. Originally the member consisted of a thick massive series of shales, some of which were black pyritiferous shales. During metamorphism the sediments were transformed into porphyroblastic orthoclase gneiss, biotite gneiss, and pyritiferous gneiss. Evidence of bedding is rare and foliation is poorly developed.

Porphyroblastic orthoclase gneiss.—Gneisses with large porphyroblasts of orthoclase are common in the west-central portion of the Dakin Hill member (Pl. 2, fig. 3). An outcrop of typical porphyroblastic orthoclase gneiss is well-exposed in the road cut 500 feet south of the junction of Highways 10 and 123 in Marlow. The porphyroblasts of orthoclase are commonly 5 cm. long and contain many inclusions. Some of the crystals are euhedral, others have been drawn out into augen. The porphyroblasts grew largely by replacement without displacing the foliation. In the center of many of the porphyroblasts, there are unreplaced portions of the gneiss. The foliation of the gneiss in the core of the porphyroblasts is parallel to the foliation in the surrounding gneiss. The percentage of porphyroblasts is highly variable, ranging from 1 to 10 per cent. In the hand specimen, the orthoclase is fairly clear and has a sub-vitreous luster. Under the microscope, a fine microperthitic structure is visible. A chemical analysis of the orthoclase from the gneiss is given in Table 9.

The matrix of the porphyroblastic orthoclase gneiss is composed of quartz, oligoclase, biotite, sillimanite, garnet, and magnetite. Muscovite does not occur in the unaltered gneisses which contain orthoclase. The variation in the amount of oligoclase is independent of the percentage of orthoclase porphyroblasts. In most specimens the oligoclase is partially replaced by quartz. Myrmekite occurs in small amounts along the borders of the orthoclase porphyroblasts. Partially altered cordierite was found in one specimen. The cordierite is laced with brown isotropic alteration material and is strongly embayed by quartz. Pale-green mica, which is probably magnesium-rich biotite, is associated with the cordierite. Modes of the gneiss are given in Table 5, columns 1 and 1a[3]; a chemical analysis is given in Table 15, column 1.

Biotite gneiss.—Biotite gneiss is common throughout most of the Dakin Hill member. Irregular lenses and clots of light-colored minerals constitute 2–20 per cent of the rock. The lenses are generally 2–5 cm. long and 1 cm. thick. Most of the light-colored aggregates are composed of muscovite and quartz with only a small amount of oligoclase-andesine. The rectangular outline of some of these aggregates suggests that they are pseudomorphs after orthoclase.

The dark portion of the gneiss is gray, medium-grained, and generally massive. The texture is granoblastic and the grain size ranges from 0.5 to 2.0 mm. In some places the gneiss has a mottled appearance due to a mixture of fine-grained and medium-grained patches about 1–2 inches in diameter; the fine-grained portions are rich in quartz and poor in garnet and biotite, whereas the medium-grained portions are rich in these dark minerals.

The biotite gneiss is composed mainly of quartz, biotite, oligoclase-andesine, sillimanite, garnet, and muscovite. Magnetite and aptite are common accessories. Modes of the gneiss are given in Table 5, columns 2 to 7. In the analyzed specimen of biotite gneiss (LM125) given in Table 15, the light and dark minerals are fairly evenly distributed throughout the rock.

Quartz generally occurs in 2–3 mm. aggregates composed of 0.5 mm. grains. Although the quartz has conspicuous strain shadows and moderately sutured borders, it is not highly granulated. Quartz forms embayments in other minerals and fills the fractures in shattered biotite grains. The quartz which penetrates plagioclase commonly has a polygonal outline. Coarse sillimanite needles generally occur in the micaceous layers and fine needles of sillimanite are conspicuous in quartz. In some of

[3] The calculation of the mode from the rock analyses is not entirely automatic in these rocks because some of the oxides occur in a large number of minerals. For this reason the quantities of a few of the minerals were determined by Rosiwal analyses (shown by an asterisk in Table 5). The mineral analyses of biotite, garnet, cordierite, and orthoclase (Table 9) were used in the calculations except for the ratio MgO:FeO in biotite. The ratio K_2O:Na_2O in muscovite was determined from the rock analysis of specimen LM358.

the quartz grains, the needles of sillimanite have a preferred orientation parallel to the long dimension of adjacent biotite flakes. Muscovite occurs in large clear flakes as well as in small grains in feldspar. In most cases the muscovite is less deformed than the biotite. Although garnets are common, they form only small crystals 1–2 mm. in diameter.

Some of the biotite gneisses in the northern portion of the Dakin Hill member contain 20–35 per cent oligoclase-andesine and have a coarse igneous-looking texture somewhat similar to that of the nonporphyritic phases of the Kinsman quartz monzonite. Although some of these gneisses occur near bodies of the Kinsman, they are also found half a mile from the Kinsman.

Pyritiferous gneiss.—Pyritiferous gneiss is common in all parts of the Dakin Hill member. The rusty-brown to jet-black weathered surface of the gneiss is distinctive in the field. In even the freshest road cuts, the gneiss is highly stained. The pyritiferous gneiss grades into biotite gneiss with which it is closely associated.

Most of the pyritiferous gneisses are massive, but in limited areas a vague foliation is developed. Lenses of light-colored minerals are not as common in the nonpyritiferous gneisses. Although the main constituents are the same as in the other gneisses in the Dakin Hill member, the pyritiferous gneiss is richer in graphite and pyrite and poorer in garnet and magnetite. Some of the gneiss contains as much as 2 per cent graphite and 5 per cent pyrite. The yellow alteration material that is present in most specimens is probably melanterite, from the decomposition of pyrite. Porphyroblasts of orthoclase are not common in the pyritiferous gneisses except in a few places in the western portion of the Dakin Hill member. Garnets normally occur only in the pyritiferous gneisses which contain orthoclase. In these rocks the garnets form large porphyroblasts up to 1 cm. in diameter, whereas garnets in the other metamorphic rocks are generally only 1–2 mm. in diameter. Modes of the pyritiferous gneiss are given in Table 5, columns 8 to 11, and a chemical analysis is given in Table 15.

Thickness

The Hubbard Hill member is about 6,000 feet thick. The beds of lime-silicate granulite and associated schists have a total maximum thickness of about 160 feet. An exact determination of the thickness of the May Pond member is not possible because data indicating the proportion of overturned beds are meager. The probable total thickness of the May Pond member is 5000 feet, but this member is absent near the northern and southern boundaries of the quadrangle. Even less information is available on the thickness of the Dakin Hill member. The great breadth of outcrop may mean that the thickness is of the order of 10,000 feet. The total maximum thickness of the Littleton formation including the 5,000 feet exposed below the Hubbard Hill member in the Bellows Falls quadrangle may be approximately 25,000 feet.

Age

No fossils have been found in the Lovewell Mountain quadrangle. However, it is possible to trace the Littleton formation through the Bellows Falls quadrangle to the Littleton-Moosilauke area. Billings and Cleaves (1934) have shown by paleontological evidence that the Littleton formation is lower Devonian in the Littleton-Moosilauke area.

Plutonic Rocks

General Statement

The plutonic rocks in the Lovewell Mountain quadrangle belong to the late Devonian (?) New Hampshire magma series and are represented by the Bethlehem gneiss, the Kinsman quartz monzonite, and the Concord granite. The Bethlehem gneiss and Kinsman quartz monzonite were metamorphosed during the closing stages of the orogeny in late Devonian time. Dikes of light-colored granitic rocks are common, but no mafic dikes have been observed. Large bodies of pegmatite are abundant in the western part of the quadrangle. Small pegmatite veins are found throughout the area but are more numerous in the plutonic rocks. The problem of the age of the New Hampshire magma series is considered after the description of the individual members.

Bethlehem Gneiss

Three bodies of Bethlehem gneiss are exposed in the Lovewell Mountain quadrangle. The

largest body occurs in the northwestern portion of the area and forms the southern extremity of the Mt. Clough pluton. A small body of Bethlehem gneiss crops out near the Sullivan-Gilsum

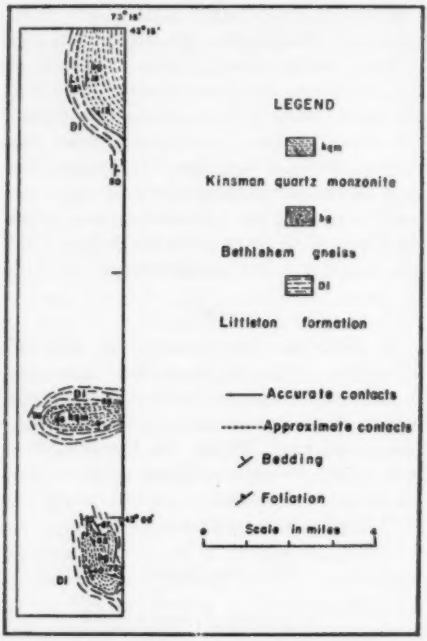

FIGURE 3.—PLUTONIC ROCKS IN THE EASTERN PART OF THE BELLOWS FALLS QUADRANGLE

town line in the southwestern part of the area. The gneiss in the west-central part of Gilsum forms the eastern prong of a body which extends for over a mile into the Bellows Falls quadrangle (Fig. 3).

The Bethlehem gneiss is light-gray to dark-gray and medium-grained to coarse-grained. Quartz augen 1 cm. long are common and large augen of feldspar as much as 2 cm. long occur locally. The Bethlehem is well foliated near the margins of the bodies, but in the central portions the foliation is poorly developed and is not as conspicuous as the lineation.

Microscopic study shows that the texture is mainly granoblastic. Cataclastic textures are found to a limited extent around some of the large feldspar augen. Essential minerals are oligoclase-andesine, quartz, microcline, biotite, and muscovite. Minor constituents are apatite, magnetite, pyrite, sphene, and zircon. The gneiss ranges from quartz monzonite to quartz diorite with granodiorite the predominant type. Modes are given in Table 6, columns 1, 2, and 3.

Most of the microcline occurs in large augen 1–2 cm. long, but some microcline is found in the groundmass. The plagioclase is not zoned and ranges in composition from An_{27} to An_{33} in different specimens. Although most of the plagioclase is fresh, it is partially sericitized in some of the gneisses near pegmatites. Quartz commonly occurs in lenses 5–10 mm. long and 1–2 mm. thick. The individual quartz grains in the lenses average 1 mm. in diameter and have not been highly granulated.

Small, partially reworked schist inclusions which range from a few inches to about a foot long are fairly common in the gneiss. A mode of one of these schist inclusions is given in Table 6, column 4. Only one large schist inclusion has been observed in the Bethlehem gneiss. This inclusion is exposed in the southwestern part of Lempster and is 1¼ miles long with a maximum thickness of half a mile. The schist is well-bedded and richer in plagioclase than the average schist of the Silver Mountain member. A mode of the schist from this inclusion is given in Table 6, column 5.

An albitized phase of the Bethlehem gneiss is exposed in the road cut 0.3 mile north of the Sullivan-Cheshire county line on Route 10. The altered rock is composed largely of albite (An_8) with some chlorite and muscovite (Table 6, column 6). Pyrite and garnet are conspicuous in some of the hand specimens. The hydrothermal solutions that caused the alteration of the gneiss may be genetically related to the pegmatites which are common in this portion of the Bethlehem gneiss.

In the Mt. Clough pluton the contacts are sharp except along the upper (eastern) contact in the northern part of the area. One of the typical sharp contacts is well-exposed in the road cut 0.4 mile south of Marlow Junction. The Bethlehem gneiss is very uniform to within a few inches of the contact. Except for moderate feldspathization near the contact, the schists are unaltered.

The northern portion of the upper (eastern) contact is somewhat gradational, especially from Dodge Pond to the northern boundary of the quadrangle. Within the Bethlehem gneiss

TABLE 6.—ESTIMATED MODES OF THE BETHLEHEM GNEISS AND ASSOCIATED SCHISTS

	Type									
	1	2	3	4	5	6	7	8	9	10
Number of thin sections	1	7	2	1	1	1	1	1	1	1
Quartz	28	30	31	59	40	2	25	26	29	30
Microcline	23	6	10	5	3			31		
Albite						83				
Oligoclase							57	30		
Oligoclase-andesine		49	45							
Andesine	36			6	44				55	46
Biotite	11	12	12	13	10		15	11	15	15
Muscovite	1	2	2	17	2	5	2	tr	1	8
Chlorite		1		tr		10	tr	tr		
Garnet		tr			1					tr
Sillimanite		tr						tr		
Myrmekite	1	tr	tr					2		1
Pyrite		tr	tr	tr		tr		tr		
Magnetite	tr	tr	tr	tr		tr	tr	tr	tr	tr
Apatite	tr	tr	tr	tr		tr	tr	tr	tr	
Zircon	tr	tr	tr	tr		tr	tr	tr	tr	tr
Sphene	tr	tr	tr				1			
Per cent of anorthite in plagioclase	32	30	32	32	30	8	28	27	30	30
Size of phenocrysts in mm.	2.0–5.0	2.0–6.0	3.0–15							
Size of groundmass in mm.	0.2–2.0	0.1–3.0	0.2–6.0	0.5–3.0	0.2–1.0	0.2–1.0	0.1–2.0	0.05–6.0	0.2–3.0	0.2–3.0

Mt. Clough Pluton.

1. Quartz monzonite.
2. Granodiorite.
3. Coarse granodiorite.
4. Schist inclusion, 0.25 mi. SW of Dodge Hollow, Lempster.
5. Schist inclusion, 0.5 mi. NE of Beaver Pond, Lempster.
6. Albitized gneiss.
7. Fine-grained gneiss from layer 2 inches thick, 1500 feet west of the upper (eastern) contact.
8. Transitional rock near the upper (eastern) contact.
9. Quartz diorite, 7′ above the base (western contact) of the pluton.

Small body at Gilsum-Sullivan town line.

10. Quartz diorite.

there are relatively fine-grained layers which range from a few inches to several feet thick. They extend across outcrops which are exposed for several tens of feet. Microscopic study of one of the fine-grained layers which occurs 1500 feet from the contact indicates that it is mineralogically similar to the normal gneiss except for a lack of potash feldspar (Table 6, column 7). Although the origin of many of these layers is not known, some of the layers near the contact represent feldspathized schist. The schists overlying the pluton have undergone considerable alteration. Some beds have been feldspathized and resemble the Bethlehem gneiss in texture and mineralogical composition; other beds have not been greatly altered. It is

not possible to decide whether some of the transitional rocks near the contact represent a contaminated phase of the Bethlehem gneiss or a granitized portion of the schist. A mode of one of the transitional rock types is given in Table 6, column 8.

The lower (western) contact of the Mt. Clough pluton is sharp and the Bethlehem gneiss near the contact appears to be homogeneous. However, microscopic study reveals significant differences in the amount of microcline in the Bethlehem. A specimen obtained 1 foot from the lower contact on Gates Mountain contained 7 per cent microcline, whereas a specimen 7 feet from the contact contained no microcline (Table 6, column 9).

The specific gravity of most specimens from the Mt. Clough pluton ranges from 2.69 to 2.72. No systematic variation in the specific gravity was observed from the base of the pluton to the upper contact.

Kinsman Quartz Monzonite

General statement.—Two bodies of Kinsman quartz monzonite occur in the Lovewell Mountain quadrangle. The body which occupies the eastern third of the quadrangle is part of a much larger pluton and is referred to in this paper as the Bacon Ledge pluton. The name Huntley Mountain pluton is applied to the smaller tabular body of Kinsman quartz monzonite in the northwestern part of the area.

Bacon Ledge pluton.—In the Bacon Ledge pluton, the Kinsman quartz monzonite is light-gray, coarse-grained, and conspicuously porphyritic. The rocks range from quartz monzonite to quartz diorite with granodiorite predominating. Modes are given in Table 7, columns 1 to 5. In calculating the modes for the porphyritic types, the percentages of phenocrysts were determined at the outcrop by linear measurements. These data were combined with the mineral percentages of the groundmass as determined in thin section to give the bulk modal composition of the rock.

The phenocrysts, composed of potash feldspar, are 3–7 cm. long, 1–3 cm. wide, and 0.5–1.5 cm. thick. They are fairly evenly distributed throughout the body and on the average comprise 10 per cent of the rock, but locally may range from 5 to 25 per cent. The phenocrysts are euhedral and are commonly oriented parallel to one another. Except in a few places where the rock has suffered late deformation, the large feldspar crystals show no signs of strain. Under the microscope the phenocrysts are seen to be microcline-microperthite. The chemical analysis given in Table 9 indicates that the feldspar is fairly low in soda with only 14 per cent of the albite molecule. Within some of the phenocrysts there are small grains of quartz, plagioclase, and biotite. These inclusions are irregularly distributed throughout the phenocrysts and generally occur in amounts less than 1 per cent.

Large garnets, 1–3 cm. in diameter, constitute 1 to 5 per cent of the rock. Although garnets are found throughout most of the pluton, their distribution at many outcrops is highly variable mainly because of the alteration of the garnet to biotite. There is a complete gradation in the alteration from fresh garnet to aggregates of biotite pseudomorphous after garnet. The chemical analysis of the garnet given in Table 9 indicates that it is largely almandite with 16 per cent of the pyrope molecule.

The groundmass of the Kinsman quartz monzonite is hypidiomorphic granular and somewhat granoblastic. Biotite-rich portions of the quartz monzonite are schistose and commonly have cataclastic textures and augen structures. The grain-size ranges from 0.05 to 8 mm. The groundmass is composed largely of quartz, oligoclase-andesine, biotite, and muscovite with minor amounts of magnetite, sillimanite, cordierite, graphite, calcite, apatite, zircon, and sphene. In only a very few rocks has potash feldspar been found in the groundmass. Normally it occurs in crystals many times larger than the other constituent minerals.

The plagioclase ranges from calcic oligoclase (An_{28}) to andesine (An_{38}) in different specimens. The plagioclase grains are 3–8 mm. in diameter and are generally larger than the other minerals in the groundmass. The plagioclase is slightly to moderately altered to sericite and kaolin with the cores more highly altered than the margins. Some blebs of potash feldspar

TABLE 7.—ESTIMATED MODES OF THE KINSMAN QUARTZ MONZONITE AND SCHIST INCLUSIONS, BACON LEDGE PLUTON

	1	2	3	3a	4	5	6	7	8	9
Number of thin sections	2	14	1		2	5	1	1	1	1
Phenocrysts and Porphyroblasts										
Microcline	13	11	14	14.0	11	11			5	
Garnet	2	2	3	3.4*		2				
Groundmass										
Quartz	30	29	28	32.0	31	29	80	26	49	35
Microcline	7	tr						25		36
Albite										25
Oligoclase									20	
Oligoclase-andesine		47				44				
Andesine	37		36	31.8	51		4	12		
Biotite	9	10	16	14.0	7	7	8	25	18	
Muscovite	2	1	3	3.0*	tr	tr	1		8	4
Chlorite		tr			tr					
Garnet					tr		3			
Cordierite						7				
Sillimanite	tr	tr	tr	0.7		tr	4	12		
Myrmekite	tr	tr	tr		tr	tr				
Graphite		tr								
Pyrite		tr	tr	0.1	tr	tr			tr	
Magnetite	tr	tr	tr	0.1	tr	tr	tr	tr	tr	tr
Ilmenite				0.3						
Apatite	tr	tr	tr	0.3	tr	tr			tr	
Zircon	tr	tr	tr		tr	tr	tr	tr	tr	tr
Sphene		tr	tr	0.3*						
Calcite		tr	tr	0.3	tr					
Tourmaline								tr		
Per cent of anorthite in plagioclase	36	35	34	34	32	35		33	28	10
Size of phenocrysts and porphyroblasts in mm	10–50	10–70	10–50		15–40	10–60			8–30	
Size of groundmass in mm	0.2–8.0	0.05–8.0	0.2–5.0		0.5–10.0	0.5–9.0	0.1–1.0	0.1–5.0	0.05–0.8	0.05–6.0

* Assumed to be equal to amount observed in thin section.

1. Porphyritic quartz monzonite.
2. Porphyritic granodiorite.
3. Porphyritic granodiorite, specimen LM1.
3a. Porphyritic granodiorite, calculated from chemical analysis, specimen LM1.
4. Porphyritic granodiorite (nongarnetiferous).
5. Cordierite bearing granodiorite.
6. Quartz-mica schist inclusion.
7. Feldspathized schist inclusion.
8. Porphyroblastic schist inclusion.
9. Altered phase of the Kinsman quartz monzonite.

are common in some of the plagioclase. A large plagioclase grain mantled by microcline was observed in one section. The rim of microcline was about 2 mm. thick and had nearly the same orientation as the core of plagioclase. No case of plagioclase mantling microcline has been found. Myrmekite is common along the borders of the microcline phenocrysts especially where the microcline is in contact with plagioclase. The plagioclase in the myrmekite is generally in optical continuity with the adjacent plagioclase in the rock. The myrmekite is undeformed and apparently grew late.

Quartz has highly sutured borders and shows undulatory extinction, but it is not greatly deformed except in the rocks which have been subjected to granulation. It has a strong tendency to embay other minerals, especially plagioclase and biotite. Most of the quartz contains many tiny needles of sillimanite. The fact that these needles are limited to quartz in most sections aids greatly in distinguishing quartz from plagioclase.

The biotite flakes are generally randomly oriented and in most specimens occur in 5 mm. aggregates. Normally the biotite is undeformed, but in the rocks which have undergone deformation, it is bent and shredded. In thin section the biotite is typically brown, but some of the biotite which has altered from garnet is green. A chemical analysis of the brown biotite is given in Table 9. In some places, biotite has altered to chlorite and muscovite. Muscovite is not as abundant as biotite and normally constitutes less than 3 per cent of the rock except in the highly altered specimens.

Cordierite is fairly common near the western contact of the pluton between South Stoddard and Camp Merriewoode. The cordierite generally comprises 3–7 per cent of the rock with local concentrations of 15 per cent not uncommon. Most of the crystals are anhedral, but some have a pseudohexagonal outline. They range in size from 8 to 20 mm. The fresh cordierite has a vitreous luster and is grayish-blue. It is normally darker than quartz, but where quartz is dark because of strain the two minerals cannot be distinguished megascopically with certainty. Most of the cordierite is highly altered to sericite.

Although sillimanite is widespread as tiny needles within quartz, it generally constitutes less than 1 per cent of the rock. Larger concentrations of sillimanite are found in the reworked inclusions. In these rocks, it normally occurs as small needles in mica, but large porphyroblasts of sillimanite are present in some of the inclusions.

Inclusions of metasedimentary rocks in the Kinsman quartz monzonite range in size from a few inches to several thousand feet. The inclusions are generally more abundant near the contacts, but the largest inclusion, which is south of Goodhue Hill in Hancock, is several miles from the contact. Inclusions in all stages of digestion may be observed. The inclusions which have been subjected to considerable reworking are mineralogically and texturally similar to the quartz monzonite but have a greater content of biotite. Modes of representative schist inclusions are given in Table 7, columns 6, 7, and 8.

Porphyroblasts of microcline and oligoclase-andesine are commonly well-developed in the inclusions. The feldspar crystals in many of the inclusions are the same size as in the quartz monzonite. Even where the porphyroblasts are small, the relative size of the microcline and the plagioclase is the same as in the enclosing rock. Included grains within the porphyroblasts of microcline are no more abundant than in the microcline of the quartz monzonite. The included biotite flakes within the porphyroblasts are randomly oriented and coarser than in the matrix of the inclusions. Evidently the biotite is not simply a relic within the porphyroblasts but has recrystallized and grown into larger flakes. There is apparently no increase in the amount of garnet in the inclusions and the size of the individual crystals is small compared to the garnets in the Kinsman quartz monzonite. Cordierite has not been found in any of the inclusions.

The matrix of some of the inclusions has an average grain size of 0.07 mm. and is thus finer-grained than any of the schist in the area. The origin of these inclusions is not clear. Their fine-grained texture is probably not due to late granulation because the feldspar porphyroblasts are largely undeformed. It is also unlikely that the grain size would have become smaller during recrystallization within the magma.

Possibly these inclusions are fragments of fine-grained schists which were brought in from outside areas.

A portion of the Kinsman quartz monzonite in the vicinity of Goodhue Hill in Antrim has been hydrothermally altered and injected by light-colored aplitic material. The highly altered rock is light-gray to white and is composed of albite, microcline-microperthite, and muscovite (Table 7, column 9). The microperthite occurs in large grains and appears to have replaced much of the albite. The plagioclase in the microperthite is albite and comprises about 25 per cent of the microperthite. It forms coarse blebs up to 1 mm. long. The muscovite occurs in aggregates composed of many tiny flakes.

Huntley Mountain pluton.—The Kinsman quartz monzonite in the southern portion of the Huntley Mountain pluton is essentially nonporphyritic and contains little potash feldspar and garnet. Quartz diorite is the predominant rock-type. Except for a slightly finer grain size, this phase is similar to the groundmass of the rocks in the Bacon Ledge pluton (Table 8, columns 1 and 2). Inclusions range in length from a few feet to several tens of feet and generally are unaltered. Small concordant bodies of Kinsman are common in the adjacent country rock.

From Tinker Pond to Huntley Mountain in Marlow, the Kinsman quartz monzonite contains relatively small phenocrysts of microcline. Generally the phenocrysts are less than 2 cm. long and comprise from 3 to 8 per cent of the rock. Garnets are uncommon and occur in small crystals. Granodiorite is the predominant rock-type but there are some phases of quartz monzonite (Table 8, columns 3 and 4). From Huntley Mountain to the northern boundary of the quadrangle, the quartz monzonite is conspicuously porphyritic and is similar in most respects to the types in the Bacon Ledge pluton (Table 8, column 5). However, no cordierite has been observed in the Huntley Mountain pluton.

In the eastern third of the Huntley Mountain pluton from Symonds Pond in Marlow to May Pond in Washington, the phenocrysts of microcline have largely altered to quartz, muscovite, and oligoclase-andesine. In some outcrops all stages in the alteration may be observed, but generally most of the microcline has been completely replaced. A mode of one of the pseudomorphs after microcline is given in Table 8, column 7. The groundmass of this rock is essentially unaltered and is similar to the groundmass of the normal porphyritic types. The alteration of the microcline apparently involved an introduction of soda and lime, but a loss of potash.

Minor bodies.—Many small bodies of Kinsman quartz monzonite occur in the metasedimentary rocks near the larger plutons, and a few bodies have been found over a mile from the main contacts. Although some of the masses are cross-cutting, most of them are concordant and are only a few tens of feet thick. They are generally composed of quartz diorite similar to the nonporphyritic phases of the Huntley Mountain pluton, but small phenocrysts of microcline occur in some types (Table 8, columns 8 and 9). The bodies in the southern part of the area have sharp contacts, whereas those in the northern portion of the May Pond and Dakin Hill members tend to have gradational contacts. In the northern area, the nonporphyritic phases of the Kinsman quartz monzonite are not greatly different from the feldspathic types of paragneiss.

Concord Granite

The body of Concord granite in the northwestern corner of the Monadnock quadrangle (Fowler-Billings, 1949) is believed to extend a short distance into the Lovewell Mountain quadrangle. Although the only outcrops in the southwestern corner of the area consist of pegmatites, the prevalent float of Concord granite suggests that a body of this granite lies below the drift.

Miscellaneous Dike Rocks

Small dikes, 1–18 inches thick, are common throughout the area, especially in the bodies of Kinsman quartz monzonite. The rocks are light gray, fine-grained to medium-grained, and range from quartz monzonite to quartz diorite. Most phases have a granitic texture but some are porphyritic with microcline phenocrysts 0.5–2 cm. long. Essential minerals are quartz,

TABLE 8.—ESTIMATED MODES OF THE KINSMAN QUARTZ MONZONITE AND MISCELLANEOUS DIKE ROCKS

	Kinsman quartz monzonite									Miscellaneous dike rocks	
	Huntley Mountain pluton							Minor bodies			
	Type										
	1	2	3	4	5	6	7	8	9	10	11
Number of thin sections	2	1	4	1	1	3	1	1	2	1	2
Phenocrysts and porphyroblasts											
Microcline			4	11	7						
Garnet					2						
Groundmass											
Quartz	38	29	33	25	21	35	40	23	28	36	36
Microcline	tr	8	5	8			3	5		32	
Oligoclase										24	53
Oligoclase-andesine			48								
Andesine	43	50		37	59	50	23	56	53		
Biotite	13	9	8	15	8	13		15	17	4	7
Muscovite	3	2	2	4	1	2	30		1	2	4
Chlorite			tr		tr						tr
Garnet	1	1	tr		2	tr			1	tr	tr
Sillimanite	2		tr		tr		3	tr	tr		
Myrmekite	tr	1	tr				1	1		2	tr
Pyrite	tr	tr			tr			tr			
Magnetite	tr	tr	tr	tr	tr	tr		tr	tr	tr	tr
Apatite	tr	tr	tr	tr	tr	tr		tr	tr	tr	tr
Zircon	tr	tr	tr	tr	tr	tr		tr	tr	tr	tr
Sphene			tr			tr					
Per cent of anorthite in plagioclase	37	34	31	37	33	35	32	34	33	23	24
Size of phenocrysts and porphyroblasts in mm			8.0–10	8.0–30	8.0–60						
Size of groundmass in mm	0.05–6.0	0.1–6.0	0.1–8.0	0.5–4.0	0.5–4.0	0.1–6.0	0.1–6.0	0.01–4.0	0.04–4.0	0.1–4.0	0.1–3.0

Kinsman quartz monzonite.
Huntley Mountain pluton.
1. Quartz diorite.
2. Nonporphyritic granodiorite.
3. Porphyritic granodiorite.
4. Porphyritic quartz monzonite.
5. Garnetiferous granodiorite.
6. Groundmass of granodiorite with altered phenocrysts.
7. Pseudomorph after microcline phenocryst.

Minor bodies.
8. Granodiorite.
9. Quartz diorite.

Miscellaneous dike rocks.
10. Quartz monzonite.
11. Quartz diorite.

oligoclase, microcline, biotite, and muscovite Modes are given in Table 8, columns 10 and 11. These rocks are somewhat similar to the Concord granite but are poorer in potash feldspar than most phases of this rock.

Age of the New Hampshire Magma Series

The Bethlehem gneiss and Kinsman quartz monzonite must be younger than lower Devonian because they cut schists of the Littleton formation. Blocks of binary granite similar to the Concord granite, which is the youngest member of the New Hampshire magma series, have been found in the Moat volcanics (Billings, 1928, p. 99–100). The Moat volcanics are considered to be Mississippian (Billings, 1941, p. 918); hence the New Hampshire magma series is middle or late Devonian. Furthermore, uraninite from pegmatites genetically related to the Bethlehem gneiss in the Cardigan quadrangle was found to be of Devonian age (Shaub, 1938, p. 339).

The age relations between the Bethlehem gneiss and the Kinsman quartz monzonite are not clear because these two members have not been observed in contact. However, the Bethlehem gneiss is slightly more deformed than the Kinsman quartz monzonite and is therefore probably older.

Mineralogy

General Statement

Many detailed studies of the minerals occurring in western New Hampshire have been made in recent years. The following discussion is mainly limited to new data on the occurrence and properties of these minerals.

Potash Feldspar

Orthoclase and orthoclase-microperthite occur as large porphyroblasts in the metamorphosed argillaceous rocks in the high-intensity portion of the high-grade zone (Fig. 8). Microcline is common in the lime-silicate rocks and is a minor constituent of the impure quartzites. These rocks crop out in the vicinity of Marlow Hill in the low-intensity portion of the high-grade zone. In the plutonic rocks, the potash feldspar is microcline and microcline-microperthite.

An analysis of orthoclase-microperthite which occurs as porphyroblasts in paragneiss (Specimen LM6) is given in Table 9, column 1. Significant data could not be obtained for the soda feldspar component because the lamellae were extremely small. The following optical data refer to the aggregate effect of the two feldspar components:

$\alpha = 1.521$ $\quad (-)2V = 70°$
$\beta = 1.525$ $\quad Z = b$
$\gamma = 1.527$ $\quad X \wedge a = 7°$

The chemical formula for the orthoclase-microperthite is as follows:

$$(K^{0.74}Na^{0.21}Ca^{0.01}Ba^{0.01})_{0.97}Al_{1.10}Si_{2.95}O_8$$

The molecular composition in weight per cent is: $Or_{77}Ab_{21}An_1Cn_1$.

An analysis of microcline-microperthite which occurs as phenocrysts in the Kinsman quartz monzonite (Specimen LM3A) is given in Table 9, column 2. In this specimen the microperthitic structure is also very fine. The following optical data, therefore, refer to the aggregate effect of the two feldspar components:

$\alpha = 1.520$ $\quad (-)2V = 80°$
$\beta = 1.523$ $\quad$ Extinction on (010) = 6°
$\gamma = 1.525$ $\quad$ Extinction on (001) = 13°

The chemical formula may be written:

$$(K^{0.80}Na^{0.14}Ca^{0.05}Ba^{0.01})_{0.97}Al_{1.08}Si_{2.96}O_8$$

The molecular composition in weight per cent is: $Or_{82}Ab_{14}An_3Cn_1$

The microcline-microperthite is fairly low in soda, but is only slightly less sodic than the average potash feldspar from the granodiorites of southern California (Larsen, 1948, p. 160). Compared to the orthoclase-microperthite from the paragneiss (LM6), the microcline-microperthite is poorer in soda. This is probably due to differences in the amount of soda available for the potash feldspar, although differences in temperature and pressure may have had some effect.

Plagioclase

The composition of the plagioclase was determined from the value of the β index of refraction. The plagioclase in all rocks, except the lime-silicate granulites and the hydrothermally

TABLE 9.—CHEMICAL ANALYSES OF MINERALS FROM THE LOVEWELL MOUNTAIN QUADRANGLE

	1	2	3	4	5	5A
SiO_2	63.90	63.94	35.34	36.97	46.87	47.33
Al_2O_3	20.46	19.48	19.10	21.66	32.57	32.59
Fe_2O_3	0.09	0.06	0.79	0.77	0.15	0.15
FeO			18.45	33.68	8.09	8.49
MnO	0.00	0.00	0.03	0.98	0.06	0.06
MgO	0.08	0.18	9.05	4.48	7.69	8.07
CaO	0.16	0.36	0.10	1.13	0.09	0.09
Na_2O	2.38	1.64	0.27	0.10	0.51	0.53
K_2O	12.61	13.48	9.15	0.08	0.93	0.48
H_2O^+	0.12	0.30	3.21	0.08	1.97	1.88
H_2O^-	n.d.	n.d.	0.17	n.d.	0.19	0.20
TiO_2	0.00	0.04	3.28	0.31	0.02	0.02
P_2O_5	n.d.	n.d.	0.03	n.d.	0.05	0.05
F	n.d.	n.d.	0.71	n.d.	n.d.	n.d.
V_2O_3	n.d.	n.d.	0.10	n.d.	n.d.	n.d.
BaO	0.36	0.39	n.d.	n.d.	n.d.	n.d.
Less O for F	—	—	-0.30	—	—	—
	100.16	99.87	99.48	100.24	99.19	99.94
S.G.	2.55	2.55	3.06	4.10	2.63	—

1. Orthoclase-microperthite from porphyroblastic orthoclase gneiss (LM6), 0.1 mile south of the junction of Highways 10 and 123 in Marlow. F. A. Gonyer, analyst.
2. Microcline-microperthite, phenocrysts in Kinsman quartz monzonite (LM3A), Franklin Pierce Highway 300 feet north of B. M. 1285, Stoddard, N.H. F.A. Gonyer, analyst.
3. Biotite from Kinsman quartz monzonite (LM1), Forest Road 300 feet east of Nelson-Antrim town-line. L.C. Peck, analyst.
4. Garnet from Kinsman quartz monzonite (LM1), Forest Road 300 feet east of Nelson-Antrim town-line. L.C. Peck, analyst.
5. Cordierite from Kinsman quartz monzonite (LM3A), Franklin Pierce Highway 300 feet north of B. M. 1285, Stoddard, N.H. L.C. Peck, analyst.

5A. Same as 5, recalculated after subtracting 4 per cent sericite.

altered rocks, is surprisingly uniform, either calcic oligoclase or sodic andesine. In the schists and paragneisses, the composition of the plagioclase ranges from An_{28} to An_{32} and averages An_{29}. The plagioclase in the lime-silicate granulites and associated rocks is very calcic (An_{87} to An_{94}). The plagioclase in the Bethlehem gneiss ranges from An_{27} to An_{33} and averages An_{31}; in the Kinsman quartz monzonite the range is from An_{30} to An_{37} and the average is An_{34}.

The average plagioclase in the middle-grade schists of the Bellows Falls quadrangle is calcic oligoclase (An_{27}) (Kruger, 1946, p. 199) and is thus similar to the plagioclase in the high-grade schists and paragneisses of the Lovewell Mountain quadrangle. This suggests that plagioclase does not become appreciably more calcic with increasing metamorphism.

Amphibole

Amphibole occurs in the lime-silicate granulites and amphibolites in the Hubbard Hill member. The amphibole in the lime-silicate granulites has a β index of 1.637 and according to Winchell's curves (1933, p. 246) it is actinolite with approximately 75 per cent of the tremolite molecule. The amphibole in the amphibolites has an average β index of 1.670 and complete optical data indicate that it is probably common hornblende.

Pyroxene

Pyroxene occurs only in the lime-silicate granulites. The β index of the pyroxene in all specimens is close to 1.690. According to Winchell's curves (1933, p. 226), the pyroxene is diopside-hedenbergite with approximately 25 per cent of the hedenbergite molecule.

Muscovite

Muscovite occurs in the schists and paragneisses except in the west-central part of the area where orthoclase is present. Much of the muscovite is late and lies across the foliation. In the plutonic rocks, it is difficult to determine what proportion of the muscovite is secondary. The muscovite that is pseudomorphic after microcline in the Kinsman quartz monzonite is obviously secondary; yet this muscovite occurs in large flakes similar to those in the groundmass. It is therefore possible that much of the muscovite in the groundmass is also secondary. The β index of the muscovite is very close to 1.598 in all rocks. From the rock analysis of Specimen LM358 (Table 15, column 3), it is

estimated that the muscovite contains about 2 per cent soda. The analyzed muscovite reported by Barth (1936, p. 780) also contains an appreciable amount of soda and is similar in optical properties to the muscovite from the Lovewell Mountain quadrangle.

Biotite

Biotite is common in nearly all the metamorphic and plutonic rocks. An analysis of biotite from the garnetiferous phase of the Kinsman quartz monzonite (LM1) is given in Table 9, column 3. The optical properties of this biotite are as follows:

$\alpha = 1.595$ $2V = 2° - 7°$; negative
$\beta = 1.650$ X = light yellowish brown
$\gamma = 1.650$ $Y = Z$ = dark reddish brown
$X < Y = Z$

The chemical formula for the biotite is:

$$K_{0.9}(Fe^{1.5}Mg^{1.0}Al^{0.4}Ti^{0.3})_{3.2}(Si^{2.7}Al^{1.3})_{4.0}O_{10}(OH^{1.6}F^{0.2})_{1.8}$$

According to Winchell's classification (Winchell, 1933, p. 267), the biotite contains 45 per cent siderophyllite, 40 per cent eastonite, 8 per cent annite, and 7 per cent phlogopite. The refractive indices of the biotite are somewhat higher than those given by Winchell (1933, p. 274) for biotite of this composition.

In the Kinsman quartz monzonite and the Bethlehem gneiss the values for the β index of the biotite generally range from 1.645 to 1.653 and average 1.647. The composition of the biotite in most of the plutonic rocks is therefore probably similar to that of the analyzed biotite.

The values for the β index of biotite in all the metasedimentary rocks except those associated with the lime-silicate granulites range from 1.635 to 1.650 and average 1.640. The biotite in these rocks apparently has a higher magnesium-iron ratio than the biotite in the plutonic rocks. The β index of the biotite in the schists interbedded with lime-silicate rocks averages 1.625. The phlogopite-eastonite molecules probably comprise 60 per cent of this biotite (Winchell, 1933, p. 274).

Garnet

Garnets are common in nearly all of the metasedimentary rocks except the pyritiferous gneisses. Apparently insufficient iron was available for the formation of garnet in most of the pyritiferous gneisses because of the large proportion of iron held in pyrite. Garnet is rare in the Bethlehem gneiss and in the Kinsman quartz monzonite in the western part of the area; however garnet comprises 1–5 per cent of the Kinsman quartz monzonite in the eastern two-thirds of the area. An analysis of garnet from the Kinsman quartz monzonite is given in Table 9, column 4. The physical properties of this garnet are as follows:

Refractive index = 1.803
Specific gravity = 4.10
Size of unit cell, a = 11.50 Å.

The chemical formula of the garnet is:

$$(Fe^{2.26}Mg^{0.54}Ca^{0.10}Mn^{0.06})_{2.96}Al_{2.04}Si_{2.97}O_{12}$$

The molecular composition in weight per cent is:

	Per cent
Almandite	81
Pyrope	16
Spessartite	2
Grossularite	1

In all of the rocks except the schists interbedded with lime-silicate granulites, the values for the refractive indices of the garnets range from 1.801 to 1.808 and average 1.805. The garnets in the metasedimentary rocks were small and contained many inclusions so that it was difficult to measure their specific gravities. Carefully selected garnet from one of the biotite gneisses was found to have a specific gravity of 4.12. The length of the edge of the unit cell was 11.50 Å. Apparently the composition of the garnets in most of the metasedimentary rocks is similar to that of the analyzed garnet in the Kinsman quartz monzonite. The garnets in the biotite schists interbedded with lime-silicate rocks have an average index of 1.788 and are therefore probably richer in the pyrope and grossularite molecules.

Cordierite

Cordierite has been observed in only two specimens of the metasedimentary rocks (Table 5, columns 1 and 9). It comprises 2 per cent or less of the rock and is highly altered to green

mica (probably magnesium-rich biotite) and light-brown isotropic matrial. Cordierite is fairly common in the Kinsman quartz monzonite near the western contact of the Bacon Ledge pluton between South Stoddard and Camp Merriewoode. Conant (1935) has described cordierite from this area, but no detailed optical or chemical studies have been made. Cordierite generally comprises 3–7 per cent of the Kinsman quartz monzonite in this zone and is moderately to highly altered to sericite (pinite). Some of the sericite occurs as fine shreds in the cordierite so that an entirely uncontaminated sample could not be obtained for chemical analysis. The analyzed cordierite which contained about 4 per cent sericite is given in Table 9, column 5. By subtracting 4 per cent sericite and correcting to 100 per cent the calculated composition of the pure cordierite given in column 5A was obtained.

TABLE 10.—OPTICALLY POSITIVE CORDIERITES

	1	2	3	4	5	6	7
SiO_2	48.37	50.15	50.09	48.19	47.96	47.69	47.33
Al_2O_3	29.22	33.07	31.78	33.45	31.56	32.52	32.59
Fe_2O_3	2.20	1.52	0.78	0.55	1.03	0.63	0.15
FeO	7.07	2.22	8.71	8.40	3.24	8.04	8.49
MnO	0.42	0.12	0.00	0.18	1.09	0.04	0.06
MgO	9.54	11.01	6.69	7.95	12.16	7.56	8.07
CaO	1.92	0.29	0.00	0.17	0.00	0.52	0.09
Na_2O	n.d.	0.14	0.00	0.22	0.33	0.53	0.53
K_2O	n.d.	0.08	0.07	0.02	tr	0.42	0.48
H_2O^+	1.84	1.37	1.43	0.67	2.80	1.85	1.88
H_2O^-	n.d.	0.09	n.d.	0.01	n.d.	0.55	0.20
TiO_2	n.d.	0.38	0.00	0.01	0.00	tr	0.02
P_2O_5	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.05
	100.58	100.44	99.55	99.82	100.17	100.35	99.94
S.G.	2.598	2.588	2.650	2.631		2.64	2.63
α	—	1.527	1.543	1.544	1.534	1.538	1.543
β	—	1.532	1.548	1.550	—	1.542	1.549
γ	—	1.538	1.553	1.556	1.543	1.547	1.555
$2V$	85°–99°	88°	88°	85°–99°	76°	84°	80°

1. From cordierite gneiss, India (Krishman, 1924) M. S. Krishman, analyst.
2. From anthophyllite gneiss, Attu (Pehrman, 1932) G. Pehrman, analyst.
3. From veined gneiss, Illmajoki (Pehrman, 1932) G. Pehrman, analyst.
4. From gneiss, Great Slave Lake area (Folinsbee, 1941b) R. E. Folinsbee, analyst.
5. From gneiss, Antarctica (Tilley, 1940) A. P. White, analyst.
6. From argillaceous hornfels, Belhelvie (Stewart, 1942) F. H. Stewart, analyst.
7. From Kinsman quartz monzonite, Lovewell Mountain quadrangle. Corrected analysis of cordierite after subtracting 4 per cent sericite. L. C. Peck, analyst.

The optical properties of the analyzed cordierite are as follows:

$\alpha = 1.543$ $\quad 2V = 80°$
$\beta = 1.549$ $\quad$ positive
$\gamma = 1.555$

The chemical formula for the cordierite may be written:

$$(Mg^{1.27}Fe^{0.74})_{2.01}Al_{4.02}Si_{4.97}O_{18}$$

This cordierite is unusual because it is optically positive. Thirteen occurrences of optically positive cordierite have been reported but only six complete analyses of optically positive

cordierite are available (Table 10). The range in composition of these cordierites is about the same as for optically negative cordierites. The most recent attempt to correlate the size of the axial angle of cordierite with chemical composition was made by Folinsbee (1941b). He obtained the curve given in Figure 4 by plotting axial angle against $K_2O + Na_2O - CaO$. The three crosses representing the cordierites which have been analyzed since Folinsbee's study fall far from his curve. The writer agrees with Stewart (1942) that sufficient data are not available to determine the relation between axial angle and chemical composition of cordierite. As in the case of the potash feldspars, it is barely possible that the thermal history of cordierite may have had some influence on the size of the axial angle.

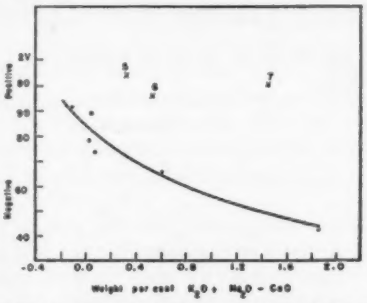

FIGURE 4.—RELATION BETWEEN AXIAL ANGLE AND $K_2O + Na_2O - CaO$ FOR CORDIERITE

Curve given by Folinsbee (1941b) based on points shown by dots. Crosses represent the cordierites which have been analyzed since Folinsbee's study. Numbers are the same as in Table 10.

Distribution of Constituents in Garnet, Biotite, and Cordierite

The analyzed garnet, biotite, and cordierite (shown by crosses in Figure 5) from the Kinsman quartz monzonite have been plotted against molecular per cent FeO, MgO, and Al_2O_3. Although the analyzed biotite and garnet were not obtained from a cordierite-bearing specimen, the optical properties of the biotite and garnet indicate that the composition of these minerals is the same as that of the garnet and biotite in the cordierite-bearing rocks. For comparison, the garnet, biotite, and cordierite from a hornfels from Belhelvie (Stewart, 1942) and garnet and cordierite from a gneiss from the Great Slave Lake area (Folinsbee, 1941a) are plotted in Figure 5. The selective distribution of FeO, MgO, and Al_2O_3 in the minerals in these rocks is about the same. The garnets are very rich in iron oxide, the cordierites are rich in magnesia, and the biotites contain nearly equal amounts of iron oxide and magnesia. It is also to be noted that although the biotites and garnets carry about the same amount of alumina, the cordierites are much richer in alumina. The distribution of MnO and TiO_2 is given in Table 11. The MnO is largely concentrated in the garnets and most of the TiO_2 occurs in the biotites.

FIGURE 5.—DISTRIBUTION OF FeO, MgO, AND Al_2O_3 IN GARNET, BIOTITE AND CORDIERITE

Crosses = minerals from the Kinsman quartz monzonite, Lovewell Mountain quadrangle; dots = minerals from hornfels, Belhelvie (Stewart, 1942); open circles = minerals from gneiss, Great Slave Lake (Folinsbee, 1941a).

TABLE 11.—DISTRIBUTION OF MnO AND TiO_2 IN GARNET, BIOTITE, AND CORDIERITE

		Garnet	Biotite	Cordierite
MnO	1.	0.98	0.03	0.06
	2.	1.81	0.02	0.04
	3.	0.66	—	0.18
TiO_2	1.	0.31	3.28	0.02
	2.	0.51	5.12	tr
	3.	0.03	—	0.01

1. Kinsman quartz monzonite, Lovewell Mountain quadrangle.
2. Hornfels, Belhelvie (Stewart, 1942).
3. Gneiss, Great Slave Lake area (Folinsbee, 1941a).

Heavy Minerals

A study of the heavy minerals from the metasedimentary and plutonic rocks was undertaken in the hope that systematic differences between the mineral assemblages in the two types of rock might be found and perhaps furnish data on the origin of these rocks. Heavy minerals were separated by heavy liquids and magnets from 14 specimens. The distribution of the minerals in weight per cent of the bulk sample is given in Table 12. Although the same heavy minerals are found in all rocks, zircon, monazite, and sphene are slightly more abundant in the plutonic rocks. The percentage of apatite reported in the plutonic rocks and schists is probably lower than it should be because some of the tiny grains in the biotite may easily have been overlooked. In the paragneisses, on the other hand, the apatite occurs in relatively large crystals and could be readily separated.

TABLE 12.—HEAVY MINERALS FROM METASEDIMENTARY AND PLUTONIC ROCKS IN WEIGHT PER CENT OF BULK SAMPLE

	Metasedimentary Rocks				Plutonic Rocks			
	1	2	3	4	5	6	7	8
Biotite	12.	27.	31.	14.	17.	16.	19.	8.
Garnet	0.3	0.8	6.	2.	0.05	2.	0.3	0.4
Sillimanite	0.0	1.	1.	4.	tr	0.03	0.0	0.03
Kyanite	0.0	tr	0.4	0.01	0.0	tr	0.0	0.0
Tourmaline	0.02	0.5	0.0	0.0	tr	0.0	0.0	0.0
Chlorite	0.0	0.0	0.0	2.	0.0	0.1	0.0	0.0
Apatite	0.03	0.05	0.8	0.1	0.2	0.1	0.03	0.03
Zircon	0.002	0.008	0.004	0.008	0.003	0.05	0.01	0.02
Monazite	0.002	0.002	0.0	0.004	0.005	0.02	0.002	0.009
Sphene	0.0	0.005	tr	0.02	0.04	0.1	0.0	0.04
Magnetite	0.06	0.07	1.	0.5	0.1	1.	0.03	0.1
Pyrite	0.0	0.0	0.0	2.	0.05	0.1	0.0	0.0
Unidentified opaques	0.3	0.3	0.5	0.2	0.2	0.3	0.2	0.3
Maximum length of zircon in mm	0.1	0.2	0.06	0.07	0.2	0.3	0.2	0.3
Maximum ratio of length to thickness of zircon	2:1	6:1	2:1	2:1	10:1	6:1	7:1	6:1

Littleton formation, middle-grade zone, Keene area.
1. Mica schist (average of 2 specimens).

Littleton formation, high-grade zone, Lovewell Mountain quadrangle.
2. Mica-quartz schist, Hubbard Hill member (average of 2 specimens).
3. Biotite gneiss, Dakin Hill member.
4. Pyritiferous gneiss, Dakin Hill member (average of 2 specimens).

Plutonic Rocks.
5. Bethlehem gneiss.
6. Kinsman quartz monzonite, Bacon Ledge pluton (average of 3 specimens).
7. Kinsman quartz monzonite, small pod in schist.
8. Granodiorite from late dikes (average of 2 specimens).

Sufficient phosphorus was available in most of the rocks for the formation of both monazite and apatite. The monazite is pale yellowish green and has a β index of 1.795. The ω index of the apatite is 1.632, indicating that it is probably fluorapatite. Tiny zircon inclusions are common in the apatite.

The zircon is colorless and its refractive indices are the same in all rocks (ω = 1.915 – 1.918). The zircons in the metasedimentary rocks are stubby, but have apparently recrystallized even in the middle-grade schists be-

cause they are euhedral and show no signs of rounding. The zircons in the plutonic rocks are larger and much more elongate. In the Kinsman quartz monzonite, the cores of some of the zircons have a "dusty" appearance, whereas the outer portions are clear. This may be due to zonal crystallization or to growth around detrital grains. One possibility is that these grains were derived from schist inclusions. However, if the Kinsman has formed by partial solution of pre-existing sediments, the zircons, in part, may be relics of these rocks.

From this study no significant differences between the heavy mineral assemblages in the various types of rocks were found except for the slightly larger amount of zircon, sphene, and monazite in the plutonic rocks.

Structural Features

General Statement

In the Lovewell Mountain quadrangle, folded metamorphosed sediments have been intruded by three members of the New Hampshire magma series. Northeasterly and northerly strikes are characteristic and dips are generally steep southeasterly or easterly except in the northwestern part of the area where dips are gentle easterly (Pl. 4). For purposes of description, the structure of the metasedimentary and plutonic rocks will be considered separately.

Metasedimentary Rocks

General Statement.—The metasedimentary rocks occur in fairly regular belts trending north-northeast in the northern part of the area and approximately north-south in the southern portion. The strata are on the east limb of a large anticline the axis of which is in the Bellows Falls quadrangle to the west; the axis of the syncline may be in the eastern part of the Lovewell Mountain quadrangle or in the area to the east. Inasmuch as the structural features of the various members of the Littleton formation are somewhat different, the individual members will be considered separately.

Hubbard Hill member.—The Hubbard Hill member forms the most westerly belt of metasedimentary rocks in the area. The Bethlehem gneiss has been emplaced within the northern portion of this member. The schists below the Bethlehem strike nearly north-south and dip very gently to the east. The schists above the Bethlehem strike consistently N 30° E and dip 55° to 80° SE. In the west-central part of the area, the Kinsman quartz monzonite cuts across the Hubbard Hill member, and the schists assume an east-west strike. The schists in the southwestern part of the quadrangle strike north-northwest and dip 55° to 75° NE.

Major folding in the Hubbard Hill member has apparently been unimportant. The lime-silicate granulites and interbedded quartz conglomerates have a simple outcrop pattern (Pl. 1), and the fact that they are exposed in only one belt indicates that there is no significant repetition of beds due to folding or faulting. Reliable data on drag folds could not be obtained in all portions of the Hubbard Hill member because of poor exposure or the vagueness of bedding. In the vicinity of Marlow Hill, however, it was possible to measure the drag folds at 17 outcrops. These data indicate that about 5 per cent of the beds face west and are consequently overturned.

Minor folds with wave lengths ranging from a few inches to several feet are common (Pl. 3, fig. 1 and 2). The axial planes of the folds are generally parallel to the schistosity and the axes are parallel to the other linear features in the schists. Contorted quartz veins are conspicuous in some places. The veins were emplaced in the later stages of deformation because they cut across the larger folds in the schists. The fact that the folds displayed by these quartz veins have the same attitude as the earlier folds in the schists indicates that the direction in which the forces acted was constant during these periods of deformation. The veins are commonly highly contorted in the beds of micaceous sillimanite schist (Pl. 3, fig. 3). and are little deformed in the beds of mica-quartz schist. Apparently the arenaceous beds were considerably more competent than the argillaceous beds.

Schistosity is well-developed throughout the Hubbard Hill member except in beds that are deficient in micaceous minerals. The most prominent schistosity is of the axial plane type. Because most folds are isoclinal, this schistosity is parallel to bedding except at the noses of

folds. Schistosites of both the axial plane and bedding type may be observed at the noses of some folds. The bedding schistosity, which is weakly developed, apparently formed before the period of tight folding and was not completely obliterated during the formation of the axial plane schistosity.

A cleavage banding has developed in some of the thin-bedded sillimanite schists at the noses of folds. The banding is parallel to the axial planes of the folds and consists of alternating 5 mm. layers of fine-grained arenaceous material and coarse micaceous material. The banding apparently results from flow of material from the beds of mica-quartz schist into the beds of sillimanite schist along the planes of schistosity.

In the Hubbard Hill member, lineation is expressed by the parallelism of prismatic minerals, streaks of minerals, elongated pebbles, and crinkles. The linear features apparently have a common origin because they are parallel to one another over fairly large areas. They are also generally parallel to the axes of the folds. In the schists east of the Bethlehem gneiss, the lineation plunges steeply southeast (Pl. 4). The lineation in the beds west of the Bethlehem plunges gently northeast. In the southwestern part of the area, the lineation plunges 40° to 65° northeast.

The stretched pebbles in the quartz conglomerates form the most striking linear feature (Pl. 3, fig. 4). The longest and intermediate axes of the pebbles lie in the plane of schistosity, whereas the shortest axis is perpendicular to the schistosity. The pebbles are elongated parallel to the axes of the minor folds. The average axial ratios of the pebbles are 1:2:8, but they range from 1:2:4 to 1:3:16. On the basis of the average axial ratios, the extension along the long axes of the pebbles is approximately 200 per cent.

May Pond member.—The May Pond member lies stratigraphically above the Hubbard Hill member. A tabular body of Kinsman quartz monzonite separates the two members in the northern half of the quadrangle. The May Pond member is composed of fairly well-banded paragneisses which show few traces of bedding. Wherever preserved, bedding is parallel to the foliation and the banding in the gneiss. The northern portion of this member strikes fairly uniformly N 30° E, but toward the south the strike is nearly east-west. Although the gneisses generally dip steeply to the east, in the vicinity of Moose and Whittemore Hills in the southwestern part of Marlow they have an average dip of 75° NW. Drag folds indicate that these beds face southeast and are thus overturned toward the southeast. This interpretation is in harmony with the theory that the younger members of the Littleton formation lie to the east.

It is difficult to determine the intensity of deformation in the May Pond member because horizon markers are lacking and bedding is obscure in most places. Generally the foliation is highly contorted and porphyroblasts of feldspar are drawn out into augen. The folds are normally small with wave lengths ranging from a few millimeters to several centimeters. The lack of cataclastic textures indicates that recrystallization accompanied or followed deformation.

Dakin Hill member.—The Dakin Hill member, which lies stratigraphically above the May Pond member, is composed of rather massive paragneisses which show practically no trace of bedding. Foliation is poorly developed, but its attitude is moderately uniform over large areas. In the southern portion of this member, the foliation strikes approximately north-south and dips steeply to the east or west. In the north, the strike is north-northeast and the dip ranges from 50° SE to vertical and averages 70° SE. The folding shown on Plate 1 is necessarily diagramatic.

Bethlehem Gneiss

Mt. Clough pluton.—The Mt. Clough pluton, named from the type locality in the Moosilauke quadrangle (Billings, 1937), is a large tabular mass of Bethlehem gneiss approximately 80 miles long and half a mile to 7 miles wide. The southern extremity of the body is exposed in the northwestern part of the Lovewell Mountain quadrangle and the northeastern part of the Bellows Falls quadrangle. The body is 6 miles wide at the northern boundaries of these quadrangles and 6 miles to the south in the Lovewell Mountain quadrangle tapers to a

point. The eastern contact is relatively straight and on the average strikes north-northeast and dips 40°–70° SE. The western contact is more sinuous and trends approximately north-south however, near the eastern contact the dip is 40°–70°E and in the northwestern part of the area some of the dips are to the west. Slab-like inclusions in various degrees of digestion are

TABLE 13.—ATTITUDE OF THE OBSERVED CONTACTS BETWEEN THE BETHLEHEM GNEISS AND SCHIST

Locality	Attitude		Exposure along strike	Height of vertical face
	Strike	Dip		
			ft.	*ft.*
Western (lower) contact				
200 ft. S60° W of the summit of Gates Mountain	N30° W	13° NE	10	6
1200 ft. S38° E of the summit of Gates Mountain	N10° E	13° SE	8	1
2150 ft. S40° E of the summit of Gates Mountain	N15° W	15° NE	20	2
2300 ft. S37° E of the summit of Gates Mountain	N15° E	15° SE	10	5
2650 ft. S35° E of the summit of Gates Mountain	N27° E	27° SE	15	3
Eastern (upper) contact				
Road cut 1900 ft. south of Marlow Jct.	N35° E	58° SE	2	3
3575 ft. S72° W of Pollards Hill	N28° E	42° SE	6	½

and dips 10°–30°E. The irregular course of the contact on Gates Mountain is due to the effects of topography on a gently dipping contact. The attitudes of the contacts as observed at seven localities are given in Table 13. The fact that the contacts of the Bethlehem gneiss appear to be approximately parallel to the bedding in the adjacent schists suggests that the body is concordant. The western contact of the body, however, is actually cross-cutting on a large scale. The base of the Mt. Clough pluton near its southern extremity is about 5000 feet above the Clough formation. Ten miles to the north, in the Claremont quadrangle, Chapman (1942, p. 914) found that the lower contact of the pluton was at the base of the Clough formation. Thus in the southern portion of the Mt. Clough pluton the cross-cutting relations are pronounced when considered over a distance of several miles.

The foliation in the Bethlehem gneiss is well-developed near the margins and is vague in the central part of the pluton. The strike of the foliation ranges from N 30° W to N 30° E and the dip generally ranges from 10° to 30° E; without exception parallel to the foliation. The fact that the foliation is sheared and the shears are filled with pegmatitic material which grades into the groundmass of the gneiss suggests that the foliation is primary. Later movements apparently took place along the foliation planes because quartz aggregates and feldspar phenocrysts are granulated and drawn out into augen parallel to the foliation. Cross-cutting pegmatites are also sheared along planes parallel to the foliation in the gneiss.

Lineation expressed by the alignment of biotite flakes and drawn out inclusions is well-developed near the eastern contact and is vague in other parts of the body. The lineation plunges 20° to 35° SSE near the contact, but decreases to approximately 10° SSE a mile from the contact (Pl. 4). The attitude of the lineation in the Bethlehem gneiss near the contact is approximately the same as in the adjacent schists. The lineation in the schists is parallel to the axes of the folds and its attitude is fairly constant for several miles from the contact of the gneiss. Inasmuch as the lineation in the schists is apparently the result of regional forces, the

lineation in the Bethlehem gneiss is probaby secondary and related to these forces rather than to magmatic movement.

Minor bodies of Bethlehem gneiss.—A small concordant body of Bethlehem gneiss, half a mile long and 400 feet wide, is exposed along the Gilsum-Sullivan town line in the southwestern part of the area. The exposure of Bethlehem gneiss west of Wright Hill in Gilsum forms the southeastern prong of a body that extends for 1 mile into the Bellows Falls quadrangle (Fig. 3). Although the contacts of this body are irregular, no cross-cutting relations were observed. The sinuosities of the contact are apparently due to interfingering of the Bethlehem gneiss and schist and to later folding.

The stratigraphic position of the bodies of Bethlehem gneiss in Gilsum suggests that they may be a part of the same sheet that formed the Mt. Clough pluton. If the projected horizon of the Mt. Clough pluton rises in the stratigraphy towards the south at the same rate as it does in the northwestern part of the area, the horizon would lie close to the bodies of Bethlehem gneiss in Gilsum. Thus these bodies may join the Mt. Clough pluton at depth or they may represent detached lenses of a once continuous sheet.

Kinsman Quartz Monzonite

Bacon Ledge pluton.—The western portion of the Bacon Ledge pluton is exposed in the eastern third of the Lovewell Mountain quadrangle. This is probably the largest body of Kinsman quartz monzonite in New Hampshire. It extends southward for 7 miles into the Monadnock quadrangle, eastward for at least 8 miles into the Hillsboro quadrangle, and northward for several tens of miles, probably into the Cardigan quadrangle.

In the Lovewell Mountain quadrangle, the general trend of the western contact is north-south and the dip ranges from 75°E to vertical. In the southern part of the area, the contact was observed at four localities and is parallel to the foliation in the adjacent paragneisses. In the northern part of the area, the actual contact was not observed, but data obtained near the contact indicates moderate interfingering of the Kinsman quartz monzonite and the paragneiss. The extension of the Kinsman into the metamorphic rocks near the village of Stoddard is apparently due to folding. Here the Kinsman quartz monzonite wraps around the nose of a

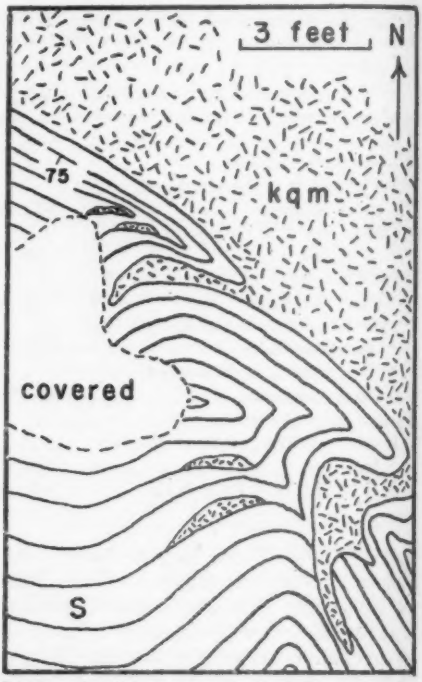

FIGURE 6.—CONTACT BETWEEN INCLUSION OF SCHIST AND KINSMAN QUARTZ MONZONITE
Pitcher Mountain, Stoddard. s = schist; kqm = Kinsman quartz monzonite.

northward-plunging anticline that has a core of paragneiss. The contacts, which were observed at five localities, are generally concordant and nearly vertical on the flanks of the anticline, but dip 55°N on the nose of the anticline (Pl. 4).

Detailed studies at favorable exposures indicated that the contacts are partly concordant and partly discordant. In some places, the Kinsman quartz monzonite is parallel to the foliation in the metasedimentary rocks; in other places it cuts across the structure (Fig. 6).

A planar structure represented by the parallel orientation of microcline phenocrysts and slab-like inclusions is fairly well-developed through-

out the pluton. Foliation due to alignment of biotite flakes is generally lacking. The fact that the large feldspar crystals are rigorously parallel to one another in many places and yet show no signs of deformation indicates that the phenocrysts were aligned during the flow stage. In the southeastern part of the area, the planar structures strike approximately north-south and most of the dips range from 75°E to vertical. In the central and northern portions of the pluton, dips as low as 20° east or west are not uncommon (Pl. 4). In this area, the planar features probably were subjected to later folding.

No distinct lineation was observed in the Kinsman quartz monzonite. The elongation of the feldspar phenocrysts is so slight that the attitude of their long axes could be determined only at the most favorable exposures. At these places, there appeared to be no preferred linear orientation.

Inclusions are most numerous along the contacts, especially where the trend of the contact is irregular. However, the largest inclusion, which is south of Goodhue Hill in Hancock, is several miles from any known contact. This inclusion contains beds of schist, quartzite, and lime-silicate granulite which are lithologically similar to the metasedimentary rocks in the Hubbard Hill member.

Huntley Mountain pluton.—The Huntley Mountain pluton lies in the northwestern portion of the quadrangle. It is a large tabular body with a minimum length of 14 miles and an average width of 1¼ miles. The body trends north-northeast in the northern portion of the quadrangle and east-west in the west-central part of the area. It extends into the Bellows Falls quadrangle for 2 miles (Fig. 3) and into the Sunapee quadrangle for an unknown distance.

The contacts are concordant except in the southern portion of the body where the Kinsman quartz monzonite commonly truncates the bedding in the schist (Fig. 7). In this part of the pluton, cross-cutting relationships on a large scale are also pronounced. Near Lake Warren in the Bellows Falls quadrangle the lower contact of the body is about 4000 feet above the base of the Littleton formation; 4 miles to the northeast, at Bald Hill in the Lovewell Mountain quadrangle, it is 10,000 feet above the base of the Littleton formation.

Angular inclusions are more common in this body than in the Bacon Ledge pluton. Possibly

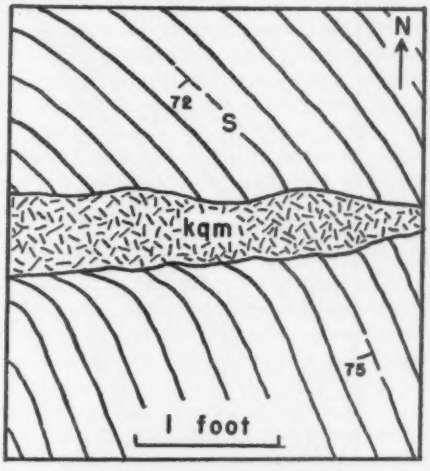

FIGURE 7.—CROSS-CUTTING STRINGER OF KINSMAN QUARTZ MONZONITE

Half a mile north of Tinker Pond, Marlow. s = schist; kqm = Kinsman quartz monzonite.

the rocks adjacent to the Huntley Mountain pluton were more brittle at the time of emplacement than were those next to the Bacon Ledge pluton. In the southern portion of the Huntley Mountain pluton, schist and quartz monzonite occur in nearly equal amounts along some of the contacts. In this area it is difficult to determine whether one is dealing with schist inclusions in the pluton or isolated masses of quartz monzonite in the schist.

Planar structures, represented by the parallelism of biotite flakes, feldspar phenocrysts, and slab-like inclusions, are well-developed except in the small apophyses which extend into the surrounding rocks. In the southern portion of the Huntley Mountain pluton, the foliation is evidently primary, because it wraps around the larger wedge-shaped inclusions. Primary movements were also important in the northern portion of the pluton because undeformed phenocrysts of feldspar show a marked parallelism. However, later movements apparently took place in this area, because a secondary folia-

tion, parallel to the planar structures in the Kinsman, is conspicuous in many of the cross-cutting pegmatite veins.

Progressive Metamorphism

In the Bellows Falls quadrangle, Kruger (1946, p. 166) has shown that there is an in-

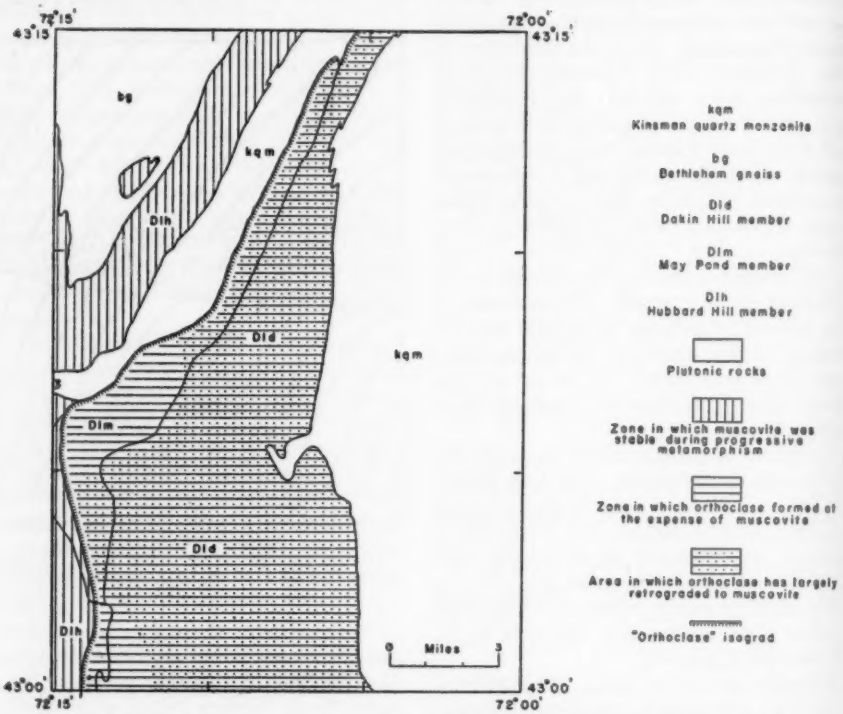

FIGURE 8.—METAMORPHIC SUBZONES IN THE LOVEWELL MOUNTAIN QUADRANGLE

METAMORPHISM OF THE METASEDIMENTARY ROCKS

General Statement

All the metasedimentary rocks and some of the plutonic rocks in the Lovewell Mountain quadrangle have been metamorphosed under high-grade conditions. In general, the intensity of the metamorphism increases from west to east. Orthoclase formed at the expense of muscovite under the highest intensity conditions; however, the orthoclase has altered to muscovite where retrograde metamorphism has occurred. Changes in chemical composition during metamorphism have not been great although locally alkalies were probably introduced or removed.

crease in the grade of metamorphism from the Connecticut River toward the east. The sillimanite isograd, which is the boundary between the middle-grade and high-grade zones, trends nearly north-south and lies about 4 miles west of the western margin of the Lovewell Mountain quadrangle. Although the high-grade zone occupies the entire Lovewell Mountain quadrangle, there is evidence that the intensity of metamorphism increases from west to east.

In the western part of the area (as shown by vertical ruling in Fig. 8) sillimanite and muscovite occur in the metasedimentary rocks and potash feldspar is lacking except in the lime-silicate rocks. East of this area, the metasedimentary rocks contain either orthoclase or aggregates of muscovite and quartz which appear to be pseudomorphous after orthoclase

(Fig. 8). The boundary between the metamorphosed argillaceous rocks which lack orthoclase and those which contain orthoclase or pseudomorphs after orthoclase is referred to as the "orthoclase" isograd. In the rocks which contain orthoclase, muscovite is lacking except where retrograde metamorphism has occurred. (See modes in Table 4, columns 1 and 2, and Table 5, columns 1 and 11.) This suggests that the orthoclase formed largely at the expense of muscovite. The fact that fine sillimanite needles commonly penetrate biotite in the gneisses east of the "orthoclase" isograd indicates that some of the potash for the orthoclase may have been derived from biotite.

Retrograde Metamorphism

In most of the metasedimentary rocks in the central portion of the quadrangle, the orthoclase porphyroblasts have been partially or completely replaced by muscovite and quartz (Fig. 8). Apparently the alteration of orthoclase occurred under high-grade conditions throughout most of this zone because sillimanite is prevalent and appears to be in equilibrium. The degree of alteration of the orthoclase increases from west to east. In the western portion of the area of retrograde metamorphism, all stages in the transition from orthoclase to aggregates of muscovite and quartz may be seen. In the partially altered porphyroblasts, muscovite occurs along cleavage planes in the orthoclase and shells of muscovite and quartz commonly surround the orthoclase. Although orthoclase is rare in the central part of the area, the form of the aggregates of muscovite and quartz indicates that they are actually pseudomorphs after orthoclase porphyroblasts.

The retrograde metamorphism was probably facilitated by solutions which were genetically related to the Kinsman quartz monzonite. Inasmuch as the pseudomorphs after orthoclase in the May Pond member contain considerable plagioclase, soda and lime were probably introduced in this zone. Throughout most of the area the orthoclase has simply been replaced by muscovite and quartz. A part of the potash in the original orthoclase was probably removed from the limits of the pseudomorphs because some of the pseudomorphs contain as little as 5 per cent muscovite. However, the occurrence of muscovite in the matrix of the gneiss indicates that the potash may have migrated only a short distance. The fact that the specimen of gneiss which has undergone retrograde metamorphism (LM 125, Table 15, column 2) contains only 0.2 per cent less potash than the specimen of orthoclase gneiss (LM6, Table 15, column 1) suggests that little if any potash was removed during the retrograde metamorphism.

In the Hubbard Hill member, sillimanite has altered to muscovite in many of the schists which are west of the "orthoclase" isograd. This alteration is not necessarily the result of retrograde metamorphism in the sense that it occurred during declining intensity conditions. Muscovite was apparently stable in these schists during the entire period of metamorphism, because orthoclase or pseudomorphs after orthoclase are lacking. Thus the alteration of sillimanite to muscovite may have occurred during progressive metamorphism as a result of the introduction of potash.

The development of chlorite in various parts of the area indicates that some retrograde metamorphism occurred under low-intensity conditions. Biotite has been most susceptible to chloritization although garnet has undergone slight alteration. Inasmuch as chlorite-bearing rocks are found sporadically throughout the area, the chloritization was probably unrelated to the alteration of the orthoclase.

Origin of the Paragneisses

The paragneisses are largely limited to the May Pond and Dakin Hill members. The chemical analyses of specimens LM6 and LM125 (Table 15, columns 1 and 2) suggest that the gneisses were originally shales. The homogeneity and lack of bedding in the gneisses indicate that the initial shales probably formed a thick series with few sandstone beds. The gneisses apparently recrystallized after most of the deformation because they are not well-foliated.

The graphite-bearing pyritiferous gneisses represented by the analyzed specimen of LM358 (Table 14, column 1) are believed to have been black shales. It is generally accepted (Twenhofel, 1939, p. 304) that the presence of carbonaceous matter and sulphur in sediments implies that there was poor circulation of the

TABLE 14.—ANALYSES OF ROCKS CONTAINING CARBON

	1	2	3	4	5	6
SiO_2	59.56	51.03	49.13	59.70	56.38	61.52
Al_2O_3	17.45	13.47	13.92	16.98	15.27	17.59
Fe_2O_3	3.10	8.06	1.35	1.31	2.81	1.54
FeO	3.84	—	1.87	4.88	3.23	4.86
MnO	0.07	—	0.25	0.16	0.09	0.04
MgO	2.67	1.15	5.11	3.23	2.84	2.76
CaO	0.21	0.78	8.73	1.27	4.23	2.04
Na_2O	0.79	0.41	0.20	1.35	1.30	1.74
K_2O	3.79	3.16	4.25	3.77	3.51	3.55
H_2O^+	2.98	—	4.69	3.82	4.09	1.96
H_2O^-	0.38	0.81	1.52	0.30	0.77	0.65
CO_2	0.00	—	6.93	1.40	3.67	0.00
TiO_2	1.10	—	0.66	0.79	0.78	0.64
ZrO_2	—	—	0.00	—	tr	—
P_2O_5	0.13	0.31	0.03	0.16	0.17	tr
SO_3	0.65	—	—	—	—	—
F	0.06	—	—	—	—	—
S	2.19	7.29	0.46*	0.63*	0.92*	0.49
C	1.20	13.11	2.00	0.46	0.59	0.87
BaO	—	—	0.00	0.08	0.08	—
Hydrocarbons	—	3.32	—	—	—	—
Total	100.17	102.90	101.10	100.29	100.73	100.25
Less O for S and F	0.86	2.73	0.18	0.24	0.34	
	99.31	100.17	100.92	100.05	100.39	

* The sulphur here given was computed from the FeS_2 reported in the original analyses. The iron in the FeS_2 was computed as Fe_2O_3. These adjustments were made so that the analyses in which FeS_2 was reported could be directly compared to the analyses in which sulphur was given separately.

1. (LM358) Pyritiferous muscovite gneiss, Lovewell Mountain quadrangle, N. H. For mode see Table 5, column 10. L. C. Peck, analyst.
2. Bituminous shale, Dry Gap, Ga. (Clarke, 1924, p. 552). L. G. Eakins, analyst.
3. Shale, Hermosa, N. M. (Clarke, 1914, p. 256). G. Steiger, analyst. 0.86 per cent pyrite reported in the original analysis.
4. Black slate, Benson, Vt. (Clarke, 1924, p. 554). W. F. Hillebrand, analyst. 1.18 per cent pyrite reported in the original analysis.
5. Black roofing slate, Slatington, Penn. (Clarke, 1914, p. 252). W. F. Hillebrand, analyst. 1.72 per cent pyrite reported in the original analysis.
6. Mica schist of the Orfordville formation, middle-grade zone, Mt. Cube area, N. H. (Hadley, 1942, p. 122). W. H. Herdsman, analyst.

water during deposition. The lack of arenaceous beds and the general massiveness of the Dakin Hill member indicates that the conditions of deposition were quiet.

Few complete analyses of black shales are available. Analyses of black slates are more common, but most of the specimens analyzed were roofing slates and are therefore probably not representative of the group. Several analyses of black shales and black slates together with the analyses of the pyritiferous gneiss from the Lovewell Mountain quadrangle are given in Table 14. The composition of the gneiss is not greatly different from that of the shales and slates except that it is richer in sulphur than most of the other rocks. However, the bituminous shale (Table 14, column 2) contains considerably more sulphur than does the gneiss; therefore it is possible that all the sulphur in the gneiss is original.

Carbonaceous rocks are not common in the low-grade and middle-grade zones of the Littleton formation, but black schists are fairly abundant in the middle-grade zone of the Orfordville formation. An analysis of one of the carbonaceous schists from the Orfordville formation in the Mt. Cube area is given in Table 14, column 6. Chapman (1939, Table 1) reported as much as 3 per cent pyrite in some of the black schists from the Orfordville formation in the Mascoma quadrangle. Thus the amount of pyrite in some of the middle-grade schists of the Orfordville formation is as great as in the high-grade gneisses of the Littleton formation.

If sulphur has been added to the pyritiferous gneisses, it probably was not introduced through simple pyritization, because the gneisses do not contain more iron than the nonpyritiferous types. Goldschmidt (1922) has suggested that carbon and sulphur may be introduced through the process of "carbon metasomation," in which introduced CS_2 or COS reacts with iron-rich silicates to form graphite, pyrite, and quartz. However, evidence that iron-silicates have been altered in the gneiss is lacking. From the general massiveness of the gneiss and its similarity in chemical composition to some of the black shales, it is concluded that the carbon and sulphur are primary constituents and that the gneiss was originally a black shale.

The light-colored nodules and bands in the gneisses are believed to have formed by replacement and segregation rather than by magmatic injection. In some gneisses, well-formed porphyroblasts of orthoclase have cores of unreplaced gneiss. Even where the orthoclase is in the form of augen, there is generally evidence that the orthoclase grew by replacement. In the gneisses containing relatively thin lenses of orthoclase, it is difficult to determine whether the lenses represent highly drawn out porphyroblasts or concentrations of a former liquid phase. In the gneisses which have not undergone retrograde metamorphism, the lenses commonly contain over 90 per cent orthoclase and less than 10 per cent quartz. One would not expect a low-temperature liquid phase to be as rich in potash. In any case, the light-colored lenses probably do not represent introduced material. The analyzed specimen of orthoclase gneiss (LM6, Table 15, column 1) actually contains slightly less potash than the analyzed slates from the low-grade zone of the Littleton

TABLE 15.—CHEMICAL ANALYSES OF METAMORPHOSED ARGILLACEOUS ROCKS FROM THE LITTLETON FORMATION

	1 LM6	2 LM124	3 LM358	4 W219	5 L489
SiO_2	62.04	58.94	59.56	58.92	58.14
Al_2O_3	18.29	18.10	17.45	18.55	21.00
Fe_2O_3*	0.33	0.49	3.10	0.94	0.33
FeO	8.45	10.04	3.84	6.63	6.32
MnO	0.22	0.25	0.07	0.08	0.06
MgO	3.23	3.45	2.67	3.24	3.41
CaO	0.45	0.89	0.21	0.48	0.32
Na_2O	0.77	1.20	0.79	1.49	1.10
K_2O	3.54	3.34	3.79	3.74	3.85
H_2O^+	0.99	1.73	2.98	3.90	4.47
H_2O^-	0.09	0.06	0.38	0.11	n.d.
CO_2	0.01	0.00	0.00	0.25	0.00
TiO_2	1.11	1.44	1.10	0.93	0.65
ZrO_2	n.d.	n.d.	n.d.	n.d.	0.04
P_2O_5	0.08	0.05	0.13	0.14	0.00
SO_3	n.d.	n.d.	0.65	n.d.	n.d.
Cl	0.00	0.01	0.00	n.d.	n.d.
F	0.10	0.10	0.06	n.d.	n.d.
S	0.03	0.15	2.19	0.17	0.09
C	n.d.	n.d.	1.20	n.d.	n.d.
Less O for S and F	−0.06	−0.12	−0.86	−0.06	—
Total	99.67	100.12	99.31	99.51	99.78
S.G.	2.85	2.88	2.72	2.75	2.777

* Includes iron in pyrite.

1. (LM6) Porphyroblastic orthoclase gneiss, Dakin Hill member, 0.1 mile south of the junction of Highways 10 and 123 in Marlow, Lovewell Mountain quadrangle. L. C. Peck, analyst.
2. (LM125) Biotite gneiss, Dakin Hill member, on road 0.5 mile west of Spruceland Camps at Granite Lake, Lovewell Mountain quadrangle. L. C. Peck, analyst.
3. (LM358) Pyritiferous muscovite gneiss, Dakin Hill member, on road 1.1 miles north of Ellisville, Lovewell Mountain quadrangle. L. C. Peck, analyst.
4. (W219) Slate, quarry at Slate Ledge, approximately 2.75 miles west of Littleton, N. H. (Billings, 1941, Table 10). G. Kahn, analyst.
5. (L489) Slate, quarry at Slate Ledge, approximately 2.75 miles west of Littleton, N. H. (Billings, 1937, p. 556). F. A. Gonyer, analyst.

formation (Table 15, columns 4 and 5). Furthermore, the amount of potash in the orthoclase gneiss is about the same as in the two analyzed gneisses which contain no potash feldspar (Specimens LM125 and LM358,

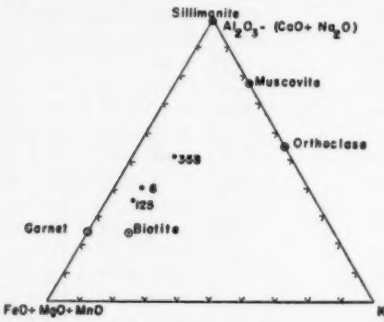

FIGURE 9.—PROJECTION OF ANALYZED PARAGNEISSES

Numbers same as in Table 15. Composition of the garnet, biotite, and orthoclase based on analyses given in Table 9. Composition of the muscovite calculated from the rock analysis of specimen LM358.

Table 15, columns 2 and 3). In Figure 9, the analyzed gneisses are plotted against molecular per cent K_2O, $Al_2O_3 - (CaO + Na_2O)$, and $FeO + MgO + MnO$. It is significant that the point representing the orthoclase gneiss (LM6) falls in the same general area in the diagram as do the points for the gneisses which lack orthoclase (LM125 and LM358). The physical conditions rather than the chemical composition have been the important factors in the formation of orthoclase in the gneisses.

The origin of the light-colored lenses in the gneisses which have undergone retrograde metamorphism is not entirely clear. In the northern portion of the May Pond member, the lenses are composed mainly of muscovite, quartz, and oligoclase. One interpretation is that the lenses represent injections of magmatic material. However, in the same area there are aggregates of muscovite, quartz, and oligoclase which appear to be pseudomorphs after orthoclase porphyroblasts, suggesting that the light-colored lenses may represent altered orthoclase lenses. Except for slight additions of lime and soda, no change in composition occurred. In the Dakin Hill member, the light-colored lenses are composed almost wholly of quartz and muscovite. Many of these lenses also probably represent altered orthoclase lenses.

Changes in Chemical Composition

General Statement.—An exact determination of the changes in composition during metamorphism was not possible because the original differences in composition between the rocks being compared were not precisely known. However, the fact that the composition of the schists and the paragneisses in the area is similar to the composition of the two analyzed slates from the Littleton formation in the Littleton-Moosilauke area (W219 and L489, Table 15) suggests that no major changes in composition occurred during metamorphism.

Hubbard Hill member.—Although no analyses of the schists from the Hubbard Hill member are available, the modes of these rocks indicate that their composition is not greatly different from that of the slates in the low-grade zone. The average schist contains 10 per cent plagioclase, about 0.8 per cent soda, and about 0.6 per cent lime. The two analyzed slates from the Littleton-Moosilauke area (W219 and L489, Table 15) average 1.30 per cent soda and 0.40 per cent lime. A few of the schists from the Hubbard Hill member contain as much as 30 per cent plagioclase. Small amounts of soda and lime may have been added to these rocks or their original composition may have been slightly different from that of the slates. If material was added, it was probably introduced by dilute solutions or vapors rather than by magmatic injection, because the light-colored constituents are evenly distributed throughout the rock. In the pseudo-sillimanite schists some potash was probably introduced during the muscovitization of the sillimanite. It is unlikely that the potash was derived from other minerals in the rock. Orthoclase is eliminated as a possible source of potash because pseudomorphs after orthoclase are lacking. Potash was not obtained from the biotite because there is no evidence that biotite has been replaced by garnet or chlorite. The fact that accessory tourmaline is fairly common in the pseudo-sillimanite schists suggests that boron may have been introduced perhaps at the same time as the potash. In the west-central portion of the

Hubbard Hill member, thin lime-silicate beds rich in lime and magnesia are closely associated with normal argillaceous rocks. Probably these compositional differences would have been eliminated if solutions had been effective in moving large amounts of material.

Some of the gneisses in the Hubbard Hill member in the southwestern part of the area contain light-colored bands rich in andesine. In general, the total amount of andesine in the gneiss is about the same as in the average schist so that the banding may be the result of segregation without addition of material. Soda and lime may have been added locally where the gneiss contains 20 to 30 per cent andesine.

May Pond and Dakin Hill members.—In the gneisses of the May Pond and Dakin Hill members, bands and nodules of light-colored minerals are common; however, the similarity in composition between the analyzed gneisses (Table 15, columns 1, 2, and 3) and the analyzed slates (Table 15, columns 4 and 5) suggests that no large changes in composition occurred during the metamorphism. In the analyzed orthoclase gneiss, Specimen LM6, orthoclase porphyroblasts comprise 8 per cent of the rock; yet the gneiss is no richer in potash than are the analyzed slates. Specimens LM6 and LM358 are slightly poorer in soda than are the slates. However, the analyzed gneisses contain only about 3 per cent plagioclase, whereas the average gneiss carries about 10 per cent plagioclase and about 1 per cent soda. Thus the soda-content of the gneiss is generally about the same as that of the low-grade rocks. Locally, where the gneiss contains as much as 35 per cent plagioclase, appreciable amounts of soda and lime may have been introduced.

Summary of Metamorphism

All the metasedimentary rocks in the area have been metamorphosed under high-grade conditions. Orthoclase developed at the expense of muscovite in the paragneisses during the peak of metamorphism. The orthoclase in the central portion of the quadrangle was largely replaced by quartz and muscovite during a period of retrograde metamorphism. Most of the metasedimentary rocks were probably derived from shales without any appreciable changes in composition during metamorphism. In portions of the Hubbard Hill member, however, potash was apparently introduced during the alteration of sillimanite to muscovite. In the few rocks which contain as much as 35 per cent plagioclase, soda and lime were presumably added.

Origin of the Plutonic Rocks

General Theory

Several lines of evidence indicate that most of the bodies of Kinsman quartz monzonite have intruded the surrounding rocks. In the case of the Bethlehem gneiss the field relations are not as conclusive, but they do demonstrate that most of the gneiss moved as a magma. Definite conclusions could not be drawn regarding the origin of the granitic material which formed these bodies. Magma rising from great depth may have entered certain favorable horizons and formed the huge sill-like masses now exposed. Some features suggest that the granitic material originated at a comparatively shallow depth by the solution or partial solution of favorable beds. The mobilized material may have moved along approximately the same stratigraphic horizon at which it originated to form the intrusive sheets.

Evidence of Intrusion

General statement.—The data pertinent to the problem of the origin of the plutonic rocks are assembled in Table 16. This material has been described in detail in the sections on plutonic rocks and structure.

Kinsman quartz monzonite.—Field evidence clearly demonstrates that most portions of the Kinsman quartz monzonite are intrusive into the surrounding rocks. In the case of some of the small nonporphyritic masses of Kinsman quartz monzonite in the northern part of the area, the data are inconclusive as to whether the rock was injected or formed essentially in place. The evidence which indicates an intrusive mode of emplacement is as follows:

1. Beds in the schist have been forced apart and disrupted by the Kinsman quartz monzonite at some of the contacts. This relation is clearly shown in the exposures at Sand Pond (Fig. 10). Many of the lenticular pods of

TABLE 16.—CHARACTERISTICS OF THE KINSMAN QUARTZ MONZONITE AND THE BETHLEHEM GNEISS

Chemistry:	Kinsman quartz monzonite—granite to quartz diorite, granodiorite predominating. Facies with large percentages of microcline phenocrysts generally granites; nonporphyritic facies generally quartz diorites. Average 4 per cent normative corundum. (For chemical analyses see Table 19.)
	Bethlehem gneiss—quartz monzonite to quartz diorite, granodioritie predominating.
Mineralogy:	Oligoclase-andesine, microcline, quartz, biotite, and muscovite; sillimanite and garnet in the bodies of Kinsman quartz monzonite in the eastern two-thirds of the quadrangle. (For modes see Tables 6, 7, 8.)
Texture:	Kinsman quartz monzonite—hypidiomorphic granular; coarse-grained, largely porphyritic with large crystals of microcline 1 to 8 cm. long. Nonporphyritic phases more common in the small bodies.
	Bethlehem gneiss—granoblastic texture; medium-grained with a few augen of microcline 1–2 cm. long.
Structure:	Bodies strikingly concordant but with some large scale cross-cutting relationships; small scale cross-cutting relationships common along many of the contacts of the Kinsman quartz monzonite.
	Dikes rare.
	Contacts generally sharp but in some places very gradational.
	Foliation fairly well-developed in the Bethlehem gneiss; poorly developed in the Kinsman quartz monzonite; phenocrysts with parallel orientation common in the Kinsman quartz monzonite. Planar structures generally parallel to contacts.
	Slab-like inclusions parallel to foliation; most inclusions in the Bethlehem gneiss considerably reworked; in the Kinsman quartz monzonite some inclusions unaltered, others highly reworked.

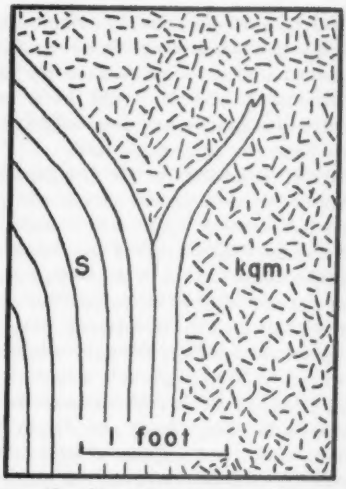

FIGURE 10.—CONTACT BETEEN SCHIST AND KINSMAN QUARTZ MONZONITE

Exposed in cliff at north end of Sand Pond, Lempster. Beds in the schist (s) have been wedged apart and disrupted by the intrusion of the Kinsman quartz monzonite (kqm.).

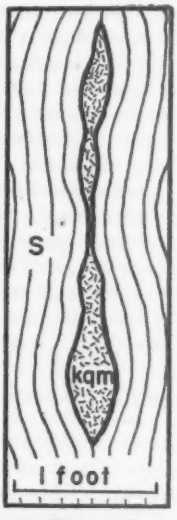

FIGURE 11.—PODS OF KINSMAN QUARTZ MONZONITE IN SCHIST

Exposed in cliff at north end of Sand Pond, Lempster. S = schist; kqm = Kinsman quartz monzonite.

Kinsman in the schists appear to be the result of intrusion. These pods are probably not boudinages because evidence of granulation and plastic deformation within the pods is lacking. Furthermore the pods are not blunt, but taper to relatively sharp points (Fig. 11).

2. Dikes of Kinsman quartz monzonite are found in the paragneisses several miles from the main plutons. A 3-foot dike is well-exposed 1 mile southwest of Big Pond in Marlow.

3. Where the Kinsman quartz monzonite cuts across the bedding of the schist inclusions,

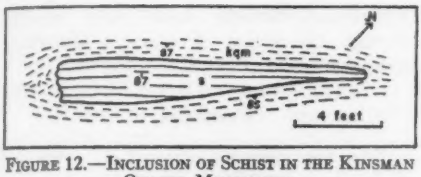

FIGURE 12.—INCLUSION OF SCHIST IN THE KINSMAN QUARTZ MONZONITE

0.3 mile west of Mack Hill, Marlow. On the southeast side of the inclusion of schist (s), the Kinsman quartz monzonite (kqm) cuts the bedding at an angle of about 10°. The foliation in the Kinsman is parallel to the contacts except at the blunt end of the inclusion where foliation is lacking.

the foliation in the Kinsman is commonly parallel to the walls of the inclusions (Fig. 12). The foliation in these cases is obviously a flow structure and not an inherited feature.

4. In the Bacon Ledge pluton, the large feldspar crystals have a parallel orientation whereas the biotite flakes are randomly oriented. If the alignment of the feldspar crystals were due to replacement along former planes of schistosity one would expect that the biotite flakes would also show a parallel orientation. The fact that the feldspar crystals are euhedral and show no signs of granulation rules out the possibility that they were oriented by differential forces acting after the solidification of the magma. The feldspars were probably oriented during the late stages of magmatic flow. Because of the large size of the feldspar crystals, they were more readily oriented than the smaller biotite flakes.

5. Although there is a gradation from quartz monzonite to schist in a few places, most of the contacts are sharp (Fig. 13). Some of the inclusions of schist near the sharp contacts, however, are highly reworked. The original composition of many of these inclusions was apparently similar to that of the wall rock. In a body which formed by replacement, one would expect the wall rock to be altered if many of the slabs of schist near the contacts are highly reworked.

6. In the middle-grade zone of the Keene quadrangle, Moore (1949) found that the Kinsman quartz monzonite was rimmed by a belt of silimanite schist which was only a few

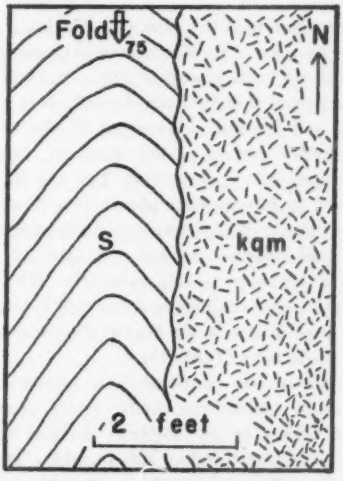

FIGURE 13.—CONTACT BETWEEN SCHIST AND KINSMAN QUARTZ MONZONITE

Half a mile northwest of Pumpkin Hill, Marlow. Bedding in schist (s) is truncated by the Kinsman quartz monzonite (kqm.).

hundred feet wide. The fact that the high-grade schists in the central part of this area are confined to narrow border zones along the pluton suggests that the Kinsman quartz monzonite intruded these rocks and furnished the heat necessary for the more intense metamorphism of the schists.

Bethlehem gneiss.—Although some of the Bethlehem gneiss may have formed in place, the following evidence indicates that large masses of the gneiss moved as a magma:

1. The fact that the schists diverge at the southern extremity of the Mt. Clough pluton (Pl. 4) indicates that they were probably intruded by the Bethlehem gneiss and wedged apart.

2. Although most of the Mt. Clough pluton is concordant, the southern portion cross-cuts the lower part of the Littleton formation. Over a distance of 10 miles, the base of the pluton rises 5000 feet in the stratigraphy.

3. Cross-cutting contacts between the Bethlehem gneiss and schist have been observed in the Mt. Cube area (Hadley, 1942, p. 165).

Evidence of Replacement

Kinsman quartz monzonite.—Evidence that large portions of the Kinsman quartz monzonite formed by replacement *in situ* is lacking. The paragneisses that have porphyroblasts of orthoclase are texturally somewhat similar to the porphyritic phases of the Kinsman, but the porphyroblastic paragneisses can not be considered as intermediate phases in a process of granitization of argillaceous rocks because their chemical composition does not approach that of the Kinsman quartz monzonite. The porphyroblastic paragneisses are as rich in alumina and as poor in soda and lime as the slates in the low-grade zone (Table 15).

Some of the nonporphyritic phases of the Kinsman quartz monzonite which occur in small bodies in the northern portions of the May Pond and Dakin Hill members are not greatly different from the paragneisses which are rich in plagioclase (Table 5, column 6). Inasmuch as cross-cutting relations have not been observed and some of the contacts are gradational, the Kinsman quartz monzonite may have originated at a comparatively shallow depth through processes analogous to those which formed the paragneiss. However, compelling evidence against an intrusive origin for these bodies is lacking.

Bethlehem gneiss.—There is no conclusive evidence that any of the Bethlehem gneiss formed in place, but some features suggest that replacement processes may have been locally important. Relatively fine-grained layers which are a few inches thick occur in the gneiss within 1500 feet of the upper (eastern) contact of the Mt. Clough pluton in the northern part of the quadrangle. Microscopic study of one of these layers indicated that it was mineralogically similar to the enclosing gneiss except for a lack of potash feldspar. It is unlikely that the fine-grained layers are younger sill-like injections because they are poorer in potash feldspar than the older rock. Another possibility is that the layers represent highly reworked inclusions. Some of the layers are only a few inches thick and are at least several tens of feet in length. One would not expect inclusions of these dimensions to remain intact if the enclosing rock moved appreciably. The fine-grained layers may represent relic beds in a portion of the gneiss which formed essentially in place. The fact that the upper contact is gradational in some places is in harmony with this idea, although the gradational contacts could have resulted through the action of solutions emanating from a magma.

Conclusions

From the evidence considered, it is concluded that at least the main portions of the Kinsman quartz monzonite and Bethlehem gneiss are intrusive into the surrounding rocks. For purposes of discussion this injected material will be referred to as magma although it may never have been entirely liquid.

Syntectonic Nature of Intrusion

The evidence in the Lovewell Mountain quadrangle supports the conclusion reached by Billings (1937, p. 537, 538) that the Bethlehem gneiss and the Kinsman quartz monzonite were intruded during the orogeny. Apparently deformation preceded the intrusion because the inclusions in the plutonic rocks are schistose. The secondary foliation that has developed in some of the pegmatites which cut the plutonic rocks indicates that deformation continued after intrusion. The Bethlehem gneiss was apparently intruded slightly earlier than the Kinsman quartz monzonite because the Bethlehem is more granulated and a secondary lineation has been imposed on it.

Metamorphism of the Plutonic Rocks

Although the Bethlehem gneiss and the Kinsman quartz monzonite were probably intruded near the close of the orogeny, they have undergone some metamorphism. In the Bethlehem gneiss, the microcline phenocrysts and large quartz grains have been drawn out into augen; however, mortar and cataclastic textures are not as common as in the Bethlehem gneiss of the Littleton-Moosilauke area (Billings, 1937, p. 506).

The Kinsman quartz monzonite in the eastern part of the area has also been affected

by the metamorphism. Although sillimanite generally comprises less than 1 per cent of the Kinsman, it is widely distributed except in the southern portion of the Huntley Mountain pluton. Some of the sillimanite needles may be magmatic or relics from incorporated schist inclusions, but the following evidence indicates that at least part of the sillimanite is metamorphic:

1. Undeformed sillimanite needles commonly extend across granulated quartz grains.

2. Sillimanite needles are concentrated in myrmekite at the margins of microcline phenocrysts which are essentially free of sillimanite.

3. The pseudomorphs after microcline in the eastern portion of the Huntley Mountain pluton contain as much as 3 per cent sillimanite. The microcline probably was altered during regional metamorphism, because the orthoclase in the nearby paragneisses of the May Pond member is similarly altered.

Garnets comprise 1–5 per cent of the Kinsman quartz monzonite in the eastern two-thirds of the quadrangle. The garnets may have originated by (1) mechanical incorporation from schist, (2) crystallization from magma, or (3) later metamorphism.

Simple mechanical incorporation of the garnets from the wall rock seems unlikely because the garnets in the Kinsman quartz monzonite are much larger than in the metasedimentary rocks.

The possibility that the garnets in the Kinsman quartz monzonite crystallized from a magma cannot be entirely ruled out; however the following evidence is against a magmatic origin:

1. In the garnetiferous phases of the Kinsman quartz monzonite, the garnets are fairly evenly distributed throughout the rock and are not concentrated in or near inclusions. If the garnets formed as a result of reaction with inclusions, one would expect them to be spatially related to the inclusions. Furthermore, reworked inclusions are common in the western part of the quadrangle; yet most of the rock is nongarnetiferous.

2. Pyrogenetic garnets in granitic rocks are generally poor in pyrope and contain appreciable amounts of spessartite (Fig. 14). The garnet from the Kinsman quartz monzonite (indicated by a cross in Fig. 14) is relatively rich in pyrope and poor in spessartite.

3. The composition of the garnetiferous Kinsman quartz monzonite is approximately the same as that of the nongarnetiferous types

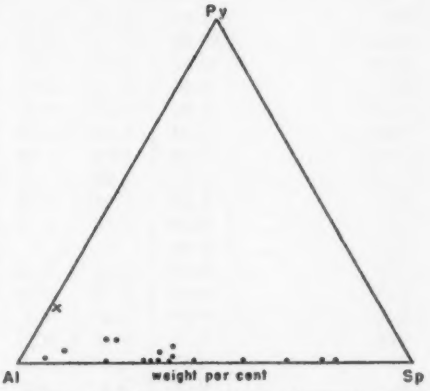

FIGURE 14.—COMPOSITION OF GARNETS FROM GRANITES

Shown by dots. Al = almandite; Sp = spessartite; Py = pyrope. Data from Wright (1938) and Hintze (1899). Cross represents the analyzed garnet from the Kinsman quartz monzonite (Table 9).

(Table 17). If the garnets are of magmatic origin, it is difficult to understand why they formed in one magma and not in another of the same composition.

The evidence which supports a metamorphic origin for the garnets is as follows:

1. As nearly as can be determined from physical properties, the garnets in the Kinsman quartz monzonite are identical in composition with those from the metasedimentary rocks in the area. Furthermore, the proportion of the major molecules in the average garnet from biotite schist in weight per cent is: 78.5% almandite, 15% pyrope, and 6.5% grossularite (Wright, 1938, p. 441). The proportion of these molecules in the garnets from the Kinsman is: 83% almandite, 16% pyrope, and 1% grossularite.

2. The garnetiferous phases of the Kinsman quartz monzonite lie in the high-intensity portion of the high-grade zone; the nongarnetiferous types generally occur in areas of less intense metamorphism.

TABLE 17.—CHEMICAL ANALYSES OF KINSMAN QUARTZ MONZONITE

	1	2	3
	Specimen No.		
	LM1	L556	L585
SiO_2	66.60	66.74	65.52
Al_2O_3	15.86	15.70	16.72
Fe_2O_3	0.35	0.85	0.60
FeO	4.48	4.30	4.75
MnO	0.06	0.06	0.06
MgO	1.50	1.50	1.55
CaO	2.64	1.96	1.94
Na_2O	2.61	2.60	2.76
K_2O	3.58	4.48	3.68
H_2O^+	0.59	0.79	1.21
H_2O^-	0.00	—	—
CO_2	0.08	0.00	0.00
TiO_2	0.95	0.70	0.70
ZrO_2	0.06	n.d.	n.d.
P_2O_5	0.23	0.11	0.23
Cl	0.01	n.d.	n.d.
F	0.08	n.d.	n.d.
S	0.14	0.10	0.19
Less O for S and F	−0.10	—	—
Total	99.72	99.89	99.91
S.G.	2.70	2.738	2.774

1. (LM1) Kinsman quartz monzonite, porphyritic granodiorite facies, Bacon Ledge pluton, Forest Road 300 feet east of Nelson-Antrim town-line, Lovewell Mountain quadrangle. L. C. Peck, analyst.
2. (L556) Kinsman quartz monzonite, porphyritic type, Kinsman Notch, outlet of Beaver Pond, Moosilauke quadrangle. (Billings, 1937, p. 556). F. A. Gonyer, analyst.
3. (L585) Kinsman quartz monzonite, nonporphyritic type, Kinsman Notch, on highway half a mile east of B. M. 1814, Moosilauke quadrangle. (Billings, 1937, p. 556). F. A. Gonyer, analyst.

For these reasons it is believed that the garnet in the Kinsman quartz monzonite formed during metamorphism, probably at the expense of biotite. No analyses of biotite from the nongarnetiferous phases of the Kinsman are available. A computation based on the rock analysis of the nongarnetiferous phases of the Kinsman quartz monzonite from the Littleton-Moosilauke area (Table 17, Specimen L585) indicates that the ratio FeO:MgO in the biotite is 1.6. In the Lovewell Mountain quadrangle, a biotite of this composition may have reacted with some of the muscovite to form the garnet (FeO:MgO = 4.2) and the biotite (FeO:MgO = 1.1) which are now found in the Kinsman.

The exact factors which caused the metamorphism of the Kinsman quartz monzonite are not known, but it is probable that moving solutions played an important part. The fact that the garnetiferous phases of the Kinsman quartz monzonite are limited to the eastern two-thirds of the quadrangle suggests that the solutions were related to the late phases of crystallization of the deeper portions of the Bacon Ledge pluton.

Origin of the Large Feldspar Crystals in the Kinsman Quartz Monzonite

The large crystals of microcline in the Kinsman quartz monzonite are probably of magmatic origin and not the result of replacement. The evidence which supports this conclusion is as follows:

1. In many places the microcline crystals show a parallel orientation; yet they are euhedral and exhibit no signs of deformation. Apparently the crystals were aligned while the magma was still partially fluid. In many places the biotite flakes in the Kinsman are randomly oriented whereas the microcline crystals show a marked parallelism. Is is therefore unlikely that the microcline crystals are porphyroblasts which grew along planes of schistosity.

2. The microcline crystals are commonly smaller in the small bodies of the Kinsman quartz monzonite. This suggests that the microcline is magmatic and that the size of the microcline crystals was influenced by the rate of cooling of the magma.

Several lines of evidence indicate that the microcline crystals are not intratelluric but grew in the later stages of the crystallization of the magma.

1. In almost every specimen, the microcline occurs only as large crystals and is not found in the groundmass. If the phenocrysts were intratelluric, a second generation of microcline would normally occur in the groundmass.

2. The fact that the phenocrysts are gener-

ally smaller in the small bodies of Kinsman quartz monzonite indicates that the growth of the crystals was probably controlled by conditions existing in the body at the time of its final emplacement.

3. Metacrysts of microcline are found in inclusions of mica-quartz schist in the Kinsman quartz monzonite. Metacrysts of potash feldspar do not occur in the mica-quartz schists outside of the plutonic bodies; therefore, the microcline must have grown after the schist had been incorporated in the magma. Presumably the inclusions were picked up at a relatively shallow depth.

4. The fact that the microcline phenocrysts contain inclusions of the other constituent minerals suggests that the formation of the microcline did not precede the growth of the other minerals.

The available data on the physical chemistry of the siliceous rocks also indicate that potash feldspar would not be the first mineral to crystallize in a quartz monzonite or granodiorite. Complete agreement regarding the position of the field boundary in the orthoclase-plagioclase equilibrium diagram is lacking (Bowen, 1928; Doggett, 1929; Vogt, 1926, 1930). The diagram proposed by Bowen (1928, p. 231) is largely hypothetical because it is based on limited experimental data. The diagram given by Doggett is based on a statistical study of fine-grained igneous rocks. Vogt (1930) made a similar study but included coarse-grained rocks. His data failed to support Doggett's diagram. The orthoclase-plagioclase equilibrium diagrams proposed by Vogt (1926, 1930) were based on the sequence of crystallization in coarse-grained plutonic rocks and extrusive rocks many of which contained phenocrysts of more than one mineral. His results are open to question because the order of crystallization in some of the rocks which he considered is not entirely clear.

From a consideration of the sequence of crystallization in fine-grained rocks, the writer has deduced the orthoclase-plagioclase diagram given in Figure 15. To eliminate errors in the interpretation of the order of crystallization as far as possible, the study was limited to those fine-grained rocks which had phenocrysts of only one mineral. The data have been obtained from the compilations of rock analyses by Clarke (1914) and by Wells (1937). For the most part, only extrusive rocks were considered; however, information on the mode of occurrence of some of the fine-grained rocks was not available, so that a number of hypabyssal rocks have probably been included. Although the entire field boundary cannot be located accurately, there is fairly good control in the more controversial portions of the diagram. The deduced boundary nearly coincides with that given by Bowen but is in poor agreement with those presented by Vogt.

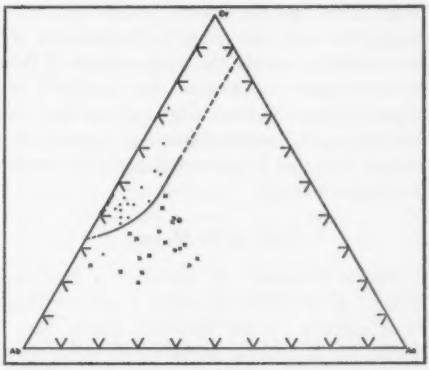

FIGURE 15.—ORTHOCLASE-PLAGIOCLASE EQUILIBRIUM DIAGRAM

Based on the order of crystallization in fine-grained igneous rocks. Dots = rocks in which orthoclase is the only phenocryst; crosses = rocks in which plagioclase is the only phenocryst; open circle = analyzed specimen (LM1) of Kinsman quartz monzonite from the Lovewell Mountain quadrangle.

The composition of Specimen LM1 is believed to be fairly representative of the Kinsman quartz monzonite in the Lovewell Mountain quadrangle. The point corresponding to this specimen (Fig. 15) falls well within the plagioclase field. It is therefore unlikely that potash feldspar would begin to crystallize before plagioclase in a magma of this composition.

Cross (1894), Pirrson (1899), Watson (1901), Pecora (1940), and Nockolds (1946) have cited cases in which the porphyritic growth of potash feldspar apparently occurred after emplacement of the magma but before final crystallization.

From the available data, it is concluded that the large microcline crystals in the Kinsman quartz monzonite are magmatic rather than the result of replacement. Most of the evidence

indicates that the large crystals are not of intratelluric origin but formed after the bodies had been partially emplaced. The reason potash feldspar tends to form large crystals in certain magmas is not clear. Most phases of the Bethlehem gneiss and the Concord granite are nonporphyritic; yet the range in composition of these rocks is mainly the same as that of the Kinsman quartz monzonite. The tendency of potash feldspar to form large crystals may be favored by high-water content of magma, although it cannot be proved that the Kinsman was richer in water.

Origin of the Magma

General statement.—As stated in a previous section, several lines of evidence indicate that most portions of the Kinsman quartz monzonite and Bethlehem gneiss are intrusive into the surrounding rocks. Regarding the origin of the magma, the following theories are considered: (1) differentiation of basaltic magma, (2) assimilation of sediments by basaltic magma, (3) "pure melting" of older granitic rocks, and (4) solution of beds at a comparatively shallow depth.

Differentiation of basaltic magma.—As Quinn (1944) pointed out, there is little evidence that the New Hampshire magma series originated by fractional crystallization of basaltic magma. The series is of restricted composition and the siliceous members are far more abundant than the basaltic ones.

Assimilation of sediments by basaltic magma.—Although inclusions of schist are common in the Kinsman quartz monzonite, it is unlikely that assimilation of schist by basaltic magma would in itself account for the observed rock types. The composition of the least amount of material which can be added to the average plateau basalt to produce the Kinsman quartz monzonite is given in Table 18, column 3. The amount of material which must be added is three times the original magma. The calculated addition does not even approximate the composition of most of the metasedimentary rocks in New Hampshire. In general the schists are richer in iron and magnesium than is the Kinsman quartz monzonite so that the simple addition of schist to a mafic magma would not produce the Kinsman. If fractional crystallization accompanied or followed assimilation of the metasedimentary rocks, some of the excess iron and magnesium might have been removed leaving a slightly aluminous magma similar to the Kinsman quartz monzonite. However, it is questionable whether sufficient heat would have been available for an appreciable amount of assimilation or whether the special conditions necessary for the postulated sequence of events would have prevailed.

TABLE 18.—COMPOSITION OF THE LEAST AMOUNT OF MATERIAL WHICH CAN BE ADDED TO AVERAGE PLATEAU BASALT TO PRODUCE THE KINSMAN QUARTZ MONZONITE

	1	2	3
SiO_2	48.80	66.60	73.2
Al_2O_3	13.98	15.86	16.8
Fe_2O_3	3.59	0.35	
FeO	9.78	4.48	2.2
MnO	0.17	0.06	—
MgO	6.70	1.50	0.0
CaO	9.38	2.64	0.6
Na_2O	2.59	2.61	2.6
K_2O	0.69	3.58	4.5
H_2O	1.80	0.59	—
TiO_2	2.19	0.95	—
P_2O_5	.33	0.23	—
S	—	0.14	—

1. Average plateau basalt (Daly, 1933, p. 17).
2. Kinsman quartz monzonite, Specimen LM1 from the Lovewell Mountain quadrangle. L. C. Peck, analyst.
3. Composition of the least amount of material which can be added to average plateau basalt to produce the Kinsman quartz monzonite.

"Pure melting" of older granitic rocks.—The fact that few members of the New Hampshire magma series are mafic suggests that the magma may have formed by the remelting of older granitic rocks. Remelting probably occurs at considerable depth so that evidence for or against this process would be difficult to obtain.

Solution of beds at a comparatively shallow depth.—The magma may have formed by processes analogous to those advocated by Billings (1941, p. 931) for the banded gneisses in the Mt. Washington area. Rocks similar in composition to the slates of the low-grade zone were believed to have been altered to gneiss by moving solutions. By selective fusion or solu-

tion of the rock, a liquid phase formed which is now represented by the light-colored streaks in the gneiss. "In the areas of most intense alteration, where 4 or 5 per cent of soda, lime, and potash were introduced, the whole rock became a molten mush similar to magma" (Billings, 1941, p. 931).

TABLE 19.—AVERAGE MODES OF BANDED GNEISS AND BETHLEHEM GNEISS

	1	2
Quartz	37	29
Plagioclase	36	45
Microcline	0	13
Biotite	15	11
Muscovite	11	2
Garnet	1	tr

1. Banded gneiss from the Mt. Washington Area (Billings, 1941, p. 880).

2. Bethlehem gneiss from the Lovewell Mountain quadrangle.

The modal composition of the gneisses formed by these processes is not greatly different from that of the Bethlehem gneiss in the Lovewell Mountain quadrangle. The average modes of the banded gneiss from the Mt. Washington area and the Bethlehem gneiss from the Lovewell Mountain quadrangle are given in Table 19. One of the most striking differences in these modes is the lack of potash feldspar in the banded gneiss. However, potash feldspar is also lacking in some phases of the Bethlehem gneiss and in nearly all the Kinsman quartz monzonite which occurs in small bodies in the metasedimentary rocks. Furthermore, a small addition of potash to the banded gneiss would convert the muscovite to potash feldspar and thus produce a rock quite similar to those phases of the New Hampshire magma series which contain microcline. Although the average banded gneiss is poorer in plagioclase than are the plutonic rocks, some of the light-colored banded gneisses contain up to 50 per cent plagioclase (Billings, 1941, p. 928). Thus in a more advanced stage of the transformation, a magma or essentially liquid mass similar in composition to the Bethlehem gneiss or the Kinsman quartz monzonite might be produced.

It may not be mere coincidence that most of the Mt. Clough pluton occurs at the same stratigraphic horizon as the banded gneiss in the Mt. Washington area. The northeast extremity of the Mt. Clough pluton lies less than 10 miles from the area of banded gneiss. Although the Mt. Clough pluton appears to be largely intrusive, the magma may have formed by the solution of beds which occurred at the same stratigraphic horizon as the pluton. Movement of the magma along this horizon would account for the intrusive features and the remarkable concordance of the body. Some of the bodies of Bethlehem gneiss which lie in areas of middle-grade metamorphism, such as the Bellows Falls pluton in the Bellows Falls quadrangle, are believed to be detached portions of the Mt. Clough pluton (Kruger, 1946, p. 190). The magma which formed these bodies probably originated at least several miles to the east. In the Lovewell Mountain quadrangle, most of the evidence also indicates that the Bethlehem gneiss has undergone movement. However, as pointed out in a previous section some features suggest that the gneiss along the upper portion of the pluton may not have moved far from its place of origin.

Conclusions.—Definite conclusions regarding the source of the magma cannot be drawn because the observed features can be explained by several different theories. Derivation of the magma from basalt by fractional crystallization or by assimilation of sediments seems unlikely, but these theories cannot be entirely ruled out. It is quite possible that the magma formed by the solution of pre-existing rocks. This process may have occurred at a relatively shallow depth because some parts of the Littleton formation have been transformed into gneiss which is chemically somewhat similar to the plutonic rocks. The solution or partial solution of certain favorable strata and movement of the mobilized material parallel to the bedding would account for the sill-like form of the bodies and the intrusive relations.

SUMMARY OF GEOLOGIC HISTORY

The earliest recorded event in the Lovewell Mountain quadrangle was the deposition of the Littleton formation during the lower Devonian. The first deposits consisted of well-stratified shales and sandy shales with a few lenses of impure dolomite, sandstone, and quartz con-

glomerate. These beds were followed by a thick series of massive shales. The total accumulation in the Lovewell Mountain and Bellows Falls quadrangles was probably 25,000 feet.

During the orogeny in the middle or late Devonian, the strata were steeply tilted to the east forming the east limb of a large anticline, the axis of which is in the Bellows Falls quadrangle. It is not known whether the axis of the syncline is in the eastern part of the Lovewell Mountain quadrangle or in the area to the east. The beds in the central part of the area were thrown into a series of large isoclinal folds. Minor folds with wave lengths ranging from a fraction of an inch to several feet also developed. Schistosities of both the bedding and axial plane types formed in the schists in the western part of the area.

Recrystallization accompanied and followed deformation. In the paragneisses of the Dakin Hill member most of the recrystallization apparently took place after deformation because foliation is poorly developed in these rocks. During the peak of metamorphism, orthoclase formed at the expense of muscovite throughout a large part of the area, but the orthoclase was replaced by muscovite and quartz during declining intensity conditions in most of the eastern two-thirds of the quadrangle. Little change in chemical composition took place during the metamorphism except for a small addition of potash to the pseudo-sillimanite schists and an addition of soda and lime to some of the paragneisses.

Near the close of the orogeny, bodies of Bethlehem gneiss and Kinsman quartz monzonite were emplaced. Definite intrusive relationships along most of the contacts indicate that the granitic material which formed these bodies moved as a magma. The granitic material may have originated at a comparatively shallow depth by the solution or partial solution of pre-existing rocks. During the last stages of the metamorphism, sillimanite and garnet formed in the Kinsman quartz monzonite in the eastern two-thirds of the area.

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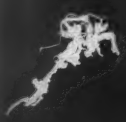

LEGEND

STRUCTURAL SYMBOLS

- Bedding
- Bedding where vertical
- Bedding where overturned
- Bedding and lineation
- Foliation (schistosity)
- Foliation (schistosity) where vertical
- Foliation (schistosity) and lineation
- Foliation (schistosity) where vertical and lineation
- Foliation and horizontal lineation
- Fold - strike and dip of axial plane and horizontal projection of axis and value of plunge
- Fold - vertical axial plane and value of plunge of axis
- Plan of folds

A-A' Ends of structure sections

FORMATION SYMBOLS

co - Concord granite

kqm - Kinsman quartz monzonite, porphyritic phase

kqmp - Kinsman quartz monzonite, with altered phenocrysts

kqmn - Kinsman quartz monzonite, non-porphyritic phase

kqma - Kinsman quartz monzonite, altered phase

bg - Bethlehem gneiss

Littleton formation:

- Dld - Dakin Hill member
- Dlm - May Pond member
- Hubbard Hill member:
  - Dlh - Mica schist
  - Dlq - Quartz conglomerates and quartzites
  - Dll - Lime-silicate granulite
  - Dlv - Amphibolite

CONTACTS

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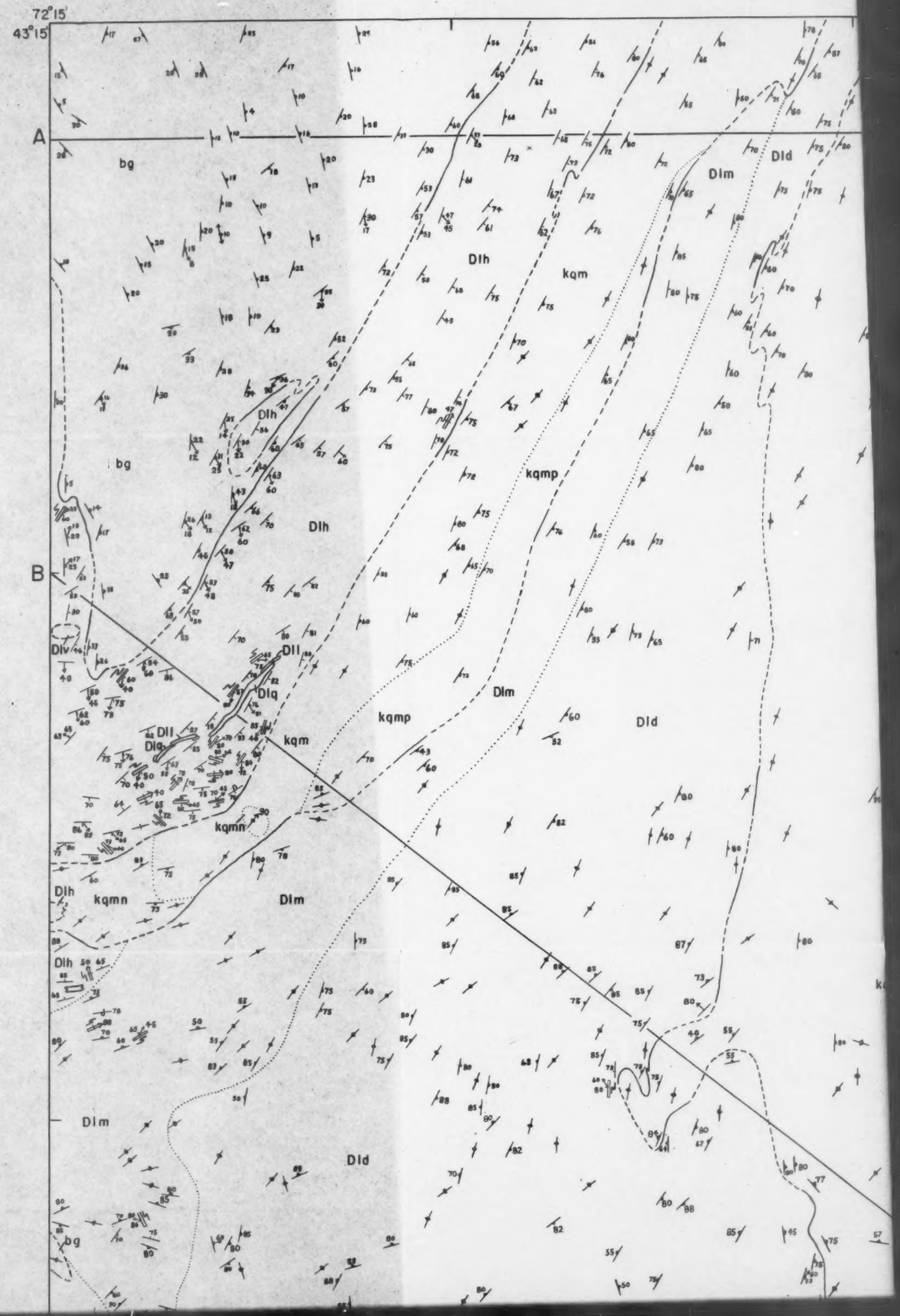

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Littleton formation

Foliation and horizontal lineation

Fold - strike and dip of axial plane and horizontal projection of axis and value of plunge

Fold - vertical axial plane and value of plunge of axis

Plan of folds

A-A' Ends of structure sections

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Dld - Dakin Hill member

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Littleton formation

Hubbard Hill member

Dlh - Mica schist

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Dll - Lime-silicate granulite

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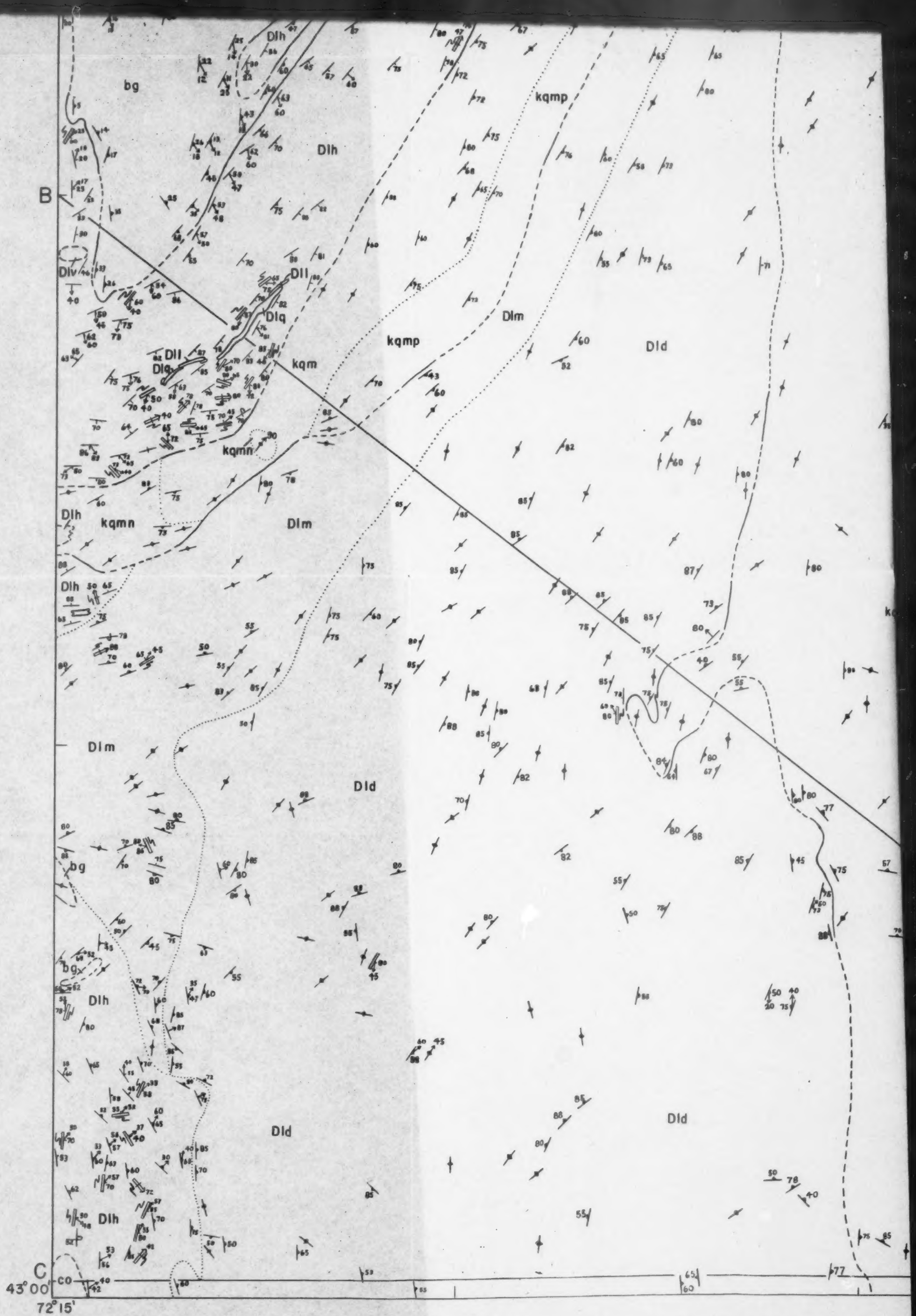

TECTONIC MAP OF THE LOVEWELL MOUNTAIN QUADRANGLE, NEW HAMPSHIRE

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TECTONIC MAP OF THE LOVEWELL MOUNTAIN QUADRANGLE, NEW HAMPSHIRE

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